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Factors Influencing Supply of Major Ions to Inland Waters, with Special Reference to the Atmosphere

Abstract: Sources of ion supply to natural inland waters include not only rocks and soils but also the atmosphere, whose significance has been underestimated. Atmospheric materials are transferred to surface waters by rain or snow, as dry fallout, or in gaseous form; the sources are the sea, land surfaces, volcanoes, products of air pollution, or organic debris. Ion supply by soil and rock weathering, which is usually more important than atmospheric supply, involves solution, oxidation-reduction reactions, activity of hydrogen ions, and complex formation. Transfer from soils to waters is influenced by ion exchange and by modes of water percolation.

Five principal environmental factors—climate, geology, topography, biota, and time—interact to determine ionic concentration and composition of atmospheric precipitation, soil solutions, and lake and river waters, although the extent to which each applies is not well understood. Investigation of situations in which only a single factor varies effectively should do much to clarify the role of each in determining the ultimate composition of natural waters.

Résumé: Les lieux d'origine des ions fournis aux eaux continentales naturelles comprennent non seulement les roches et les sols, mais aussi l'atmosphére, dont le rôle a été jusqu'à présent sousestimé. Les matériaux atmosphériques sont transportés jusqu'aux eaux de surface par la pluie et la neige, en pluie de poussières, ou à l'état gazeux; ils peuvent provenir de la mer, des surfaces terrestres, des volcans, des impuretés atmosphériques, et de débris organiques. L'apport d'ions par la décomposition des sols et des roches, qui est en général plus important que l'apport atmosphérique, nécessite la dissolution, des réactions d'oxidoréduction, l'activité d'ions-H et la chélation. Le transfert depuis les sols jusqu'à l'eau est sous l'influence des échanges d'ions et du mode de percolation de l'eau.

Cinq facteurs principaux du milieu—climate, géologie, topographie, biota et temps—réagissent les uns sur les autres pour déterminer la concentration ionique et la composition des apports atmosphériques, des solutions des sols, et des eaux lacustres et fluviales, bien qu'on ne comprenne pas encore bien le rôle de chacun d'entre eux. L'étude de situations dans lesquelles un seul de ces facteurs varie effectivement devrait permettre d'élucider le rôle de chacun d'eux dans la composition ultime des eaux naturelles.

Resumen: Las fuentes suministradoras de iones de las aguas continentales naturales no comprenden solamente las rocas y suelos, sino también la atmósfera, cuya importancia ha sido subestimada hasta ahora. Los materiales atmosféricos son transferidos a las aguas superficiales por la lluvia o nieve, a través de la caída en seco o en forma gaseosa, siendo las fuentes el mar, las superficies, los volcanes, las poluciones del aire y los deshechos orgánicos. El suministro de iones por los suelos y por el intemperismo de las rocas, el cual es generalmente de más importancia que el suministro atmosférico, implica la disolución, las reacciones de reducción y oxidación, la actividad de iones de hidrógeno y la descomposición por reacciones orgánicas. El traspaso desde los suelos a las aguas está influenciado por el intercambio de iones y por las modalidades de la percolación del agua.

Cinco factores principales del medio ambiente—clima, geología, topografía, biota y tiempo—actúan entre sí para determinar la concentración iónica, la composición de la precipitación atmosférica, de las soluciones en el suelo y de las aguas de los lagos y ríos, aunque no se sabe bien la importancia que tiene cada factor. La investigación de las situaciones en las que cada uno de los factores varía efectivamente, hará mucho para aclarar el papel que tiene cada uno en determinar la composición fundamental de las aguas naturales.

Zusammenfassung: Zu den Quellen, durch die den natürlichen Binnengewässern Ionen zugeführt werden, gehören nicht nur Gesteine und Erden, sondern auch die Atmosphäre, deren Bedeutung bisher unterschätzt worden ist. Atmosphärische Stoffe werden den offenen Gewässern durch Regen oder Schnee, als trockener Niederschlag oder in Form von Gas zugeführt, wobei die Quellen das Meer, Landoberflächen, Vulkane, Luftverunreinigungen oder organische Überbleibsel sind. Ionenzu-

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fuhr durch Verwitterung von Erden und Gesteinen, die gewöhnlich wichtiger ist als atmosphärische Zufuhr, schliesst Auflösung, Oxidation-Reduktionserscheinungen, Wirksamkeit von Wasserstoff-Ionen und Chelation ein. Die Verlagerung von Erden in Gewässer wird durch Ionen-Austausch und durch die Art, wie das Wasser durchgesickert ist, beeinflusst.

Fünf prinzipielle Umgebungsfaktoren—Klima, Geologie, Topographie, Lebensraum und Zeit—beeinflussen sich gegenseitig und bestimmen die Konzentration der Ionen und die Zusammensetzung von atmosphärischen Niederschlägen, mineralischen Lösungen, sowie die des Wassers von Seen und Flüssen. Es ist jedoch nicht ganz klar, bis zu welchem Grade jeder der Faktoren daran beteiligt ist. Die Untersuchung von Situationen, in denen nur einzelne Faktoren wirkungsvoll variieren, sollte viel dazu beitragen, die Rolle jedes einzelnen bei der Bestimmung der tatsächlichen Zusammensetzung der natürlichen Gewässer zu klären.

ЭВИЛЛ ГОРХЭМ

Факторы, Влияющие на Приток Основных Ионов во Внутренние Водоемы и Роль Атмосферы

Резюме: Источником ионов в водах внутренних водоемов являются не только горные породы и почвы, но и атмосфера, значение которой недооценивалось до сего времени. Атмосферное вещество попадает в поверхностные воды с дождем или снегом, в виде сухого осадка, либо в газообразной форме. Источником этого материала являются моря, земная поверхность, загрязнение атмосферы, и органические остатки. Почва и выветривание горных пород обычно являются более важными источниками ионов нежели атмосфера. Этот процесс заключается в растворении, окислительно-восстановительных реакциях, действии водородных ионов и в образовании комплексных соединений. Перенос ионов из почвы в водоемы зависит от ионного обмена и от условий циркуляции воды.

Климатические условия, геологический характер местности, топографические условия, биом, и время являются пятью главными факторами, определяющими характер среды и обуславливающими концентрацию ионов, состав атмосферных осадков, почвенных растворов, озерных и речных вод. Нужно признать, однако, что пределы влияния каждого фактора на характер среды не вполне ясны. Исследование условий, при которых изменяются лишь единичные факторы, приведут к выяснению роли каждого фактора в отдельности. Это может быть произведено путем полного анализа состава природных вод.

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INTRODUCTION

Although the fresh waters of the world are infinitely varied in their ionic composition, four cations—calcium, magnesium, sodium, and potassium—and three anions—bicarbonate, sulfate, and chloride—account for all but a small fraction of the total ionic concentration. The general proportions of these major cations and anions are given in Table 1, following computations of Clarke (1924, p. 119) and Conway (1942, p. 130).

It is of course most unlikely that ionic proportions in natural waters should reflect the make-up of the earth's crust, even with respect to the major ions mentioned. A brief consideration of Finnish river waters and rocks illustrates this point. Finnish bedrocks, like most igneous rocks, are very deficient in the carbonates, sulfates, and chlorides, which bulk so large in Finnish and other natural waters. A comparison of the relative proportions of alkalies and alkaline earths in the mainly granitic bedrock of Finland with their proportions in Finnish river waters (Viro, 1953, p. 15) indicates marked differences in cations (Table 2). The proportion of calcium in these Finnish river waters is nearly half as much again as in the rocks, while potassium in the waters is only about half the proportion found in the rocks. The Finnish river waters are also proportionately richer in magnesium and poorer in sodium than is the igneous bedrock. It thus

TABLE 1. IONIC COMPOSITION OF THE AVERAGE FRESH WATER OF THE WORLD
(Conway, 1942)
Total ionic concentration 2.35 m equiv./l

	m equiv. %
Cations	
Ca	64
Mg	17
Na	16
K	3
Anions	
HCO_3	73
SO_4	16
Cl	10

appears that the alkaline earths calcium and magnesium are dissolved from the soils of Finland to a greater extent than are the alkalies sodium and potassium, although in general one might expect the opposite relationship (Polynov, 1951). In these calculations no account has been taken of the ions supplied by rain, but data are available for two of the major Finnish rivers (Table 4), and the cation proportions of river water from which the rain contribution has been subtracted are also given in Table 2. The conclusions remain unchanged, and it seems likely that most of the easily soluble alkalies have been leached from Finnish soils, allowing alkaline earths to predominate. It would be interesting to know how general this phenomenon is among inland waters on substrata of igneous origin.

TABLE 2. EQUIVALENT PROPORTIONS OF ALKALIES AND ALKALINE EARTHS IN THE BEDROCK AND MAJOR RIVER WATERS OF FINLAND
(Calculated from Viro, 1953)

	Ca	Mg	Na	K
	m equiv. % of total alkalies plus alkaline earths			
Bedrock	32	22	26	20
Five rivers	45	27	19	9
Two rivers minus rain contribution	41	28	21	11

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PREVIOUS INTERPRETATIONS OF THE CHEMISTRY OF INLAND WATERS

Climatic Influences

Despite wide variations in the chemistry of lake and river waters, a number of attempts have been made to establish general rules governing their composition. Clarke (1924, p. 93) early pointed out that, in more humid and fertile regions where vegetation is abundant, bicarbonate is commonly the dominant anion in river waters, whereas sulfate and chloride commonly predominate in the more concentrated river waters of arid and unproductive areas. Table 3 compares the Hudson River in New York, draining moist and fertile areas, with the Rio Grande in New Mexico, draining arid regions (Clarke, 1924, p. 76 and 87). Bicarbonate is by far the most abundant anion in the Hudson water, while in the concentrated water of the Rio Grande it is outweighed by the sum of sulfate and chloride. Among the cations calcium predominates strongly in the Hudson water, while in the Rio Grande sodium is nearly as important. Under humid tropical conditions surface waters are very poor in dissolved salts but are also characterized by high proportions of silica. An analysis for the extremely dilute water of the Essequibo River in British Guiana (Clarke, 1924, p. 95) is included in Table 3. Since the waters are brown and rich in organic matter, this river probably drains tropical podzols rather than lateritic soils (*see* Richards, 1941).

Geologic Influences

Clarke (1924, p. 94) also remarked that, although climate is of major importance, local geological variation may greatly affect the character of rivers, especially near their sources. In large rivers such variations are evened out by the intermixing of tributaries, which "tends to produce an average composition which may be called that of a normal water."

The influence of geology was treated by Clarke mainly in its local aspects. Its more general significance has been considered by Conway (1942, p. 134), who examined the way in which concentrations and proportions of major ions change in relation to increases in the total salinity of river waters, using analyses compiled by Clarke.

According to Conway, waters of a salinity less than 50 ppm represent drainage from

igneous and metamorphic rocks, since their average proportions agree well with those of Bohemian rivers known to flow through crystalline rocks only. Up to this level of salinity the major ions rise more or less in proportion to salinity. Above this level, calcium, magnesium, and bicarbonate increase rapidly until they reach more or less constant concentrations at about 200 ppm total salinity; at the same time sodium, potassium, and chloride show relatively little change. In this range Conway believes that drainage from sedimentary rocks becomes more and more important as salinity rises. Above about 200 ppm salinity, sodium, potassium, and chloride undergo a second rapid rise in concentration, while calcium, magnesium, and bicarbonate remain constant; this is interpreted as due to drainage from arid saline soils or salt beds, and possibly (although improbably) to large-scale pollution.

Table 3. Comparison of Ionic Composition in Waters of Arid and Humid Regions
(Data from Clarke, 1924)

	Ca	Mg	Na	K	HCO_3 and CO_3	SO_4	Cl	Total salts
	m equiv.% of total cations or anions							m equiv./l
Rio Grande at Mesilla, New Mexico	47	11	40	3	36	41	24	6.37
Hudson River at Hudson, New York	62	18	17	3	73	20	7	1.78
Essequibo River above Wataputa Falls, British Guiana	26	32	40	2	69	16	13	0.23

Chemistry of Lakes in Closed Basins

Lake waters in closed drainage basins without outlets to the ocean owe their high salinity primarily to climatic aridity. The median concentration appears to be in the vicinity of 10,000 ppm (Hutchinson, 1957, p. 565). According to Langbein (In press) the universal factor determining the existence of closed lakes is evaporation, which varies with temperature and precipitation. The climatological limits for such lakes should be set by an excess of precipitation over evaporation from the lake, and the following values (from Langbein's Fig. 3) are given for the bounding climatic conditions:

Temperature (° F)	40	50	60
Precipitation (inches)	26	32	39

Thus annual temperatures higher than 40° F would theoretically permit development of a closed lake where precipitation equals 26 inches, whereas precipitation greater than 39 inches in an area of 60° F mean annual temperature would prevent such development. Langbein points out that closed lakes do not occur as widely as the above limits allow, since these presuppose absence of tributaries whose inflow would cause overflow from the lake. Moreover, near the theoretical climatic boundary, annual fluctuations in precipitation would lead to lakes being closed one year and open another, so that they would not be recognized as true closed lakes. These two factors therefore restrict the distribution of closed lakes to regions where evaporation is appreciably in excess of precipitation.

Langbein also remarks that most closed lakes are far from saturated with respect to one or more of the common salts in natural waters, suggesting that accumulation must be offset by some form of wastage, as for example by occasional overflow or by deflation or burial of salts by sediment following a period of desiccation. He further suggests that the salts dissolved in a closed lake will generally reach equilibrium under a stable climate, but will fluctuate over long periods as the balance between input and loss alters. Equilibrium salinity is shown to depend not only upon climate but also upon the shape of the basin.

Clarke (1924, p. 180) and Hutchinson (1957, p. 566) set forth the changes to be expected in ionic proportions as a normal fresh water is concentrated by evaporation in a closed basin. First, the relatively insoluble carbonates of the alkaline earths will precipitate, followed by more soluble calcium sulfate.

Then sodium carbonate and sulfate will precipitate, and later magnesium sulfate and sodium chloride, leaving the extremely soluble chlorides of the alkaline earths. The theoretical sequence is thus from carbonate to sulfate-chloride to chloride waters. However, if the waters are initially low in alkaline earths, these may be almost wholly precipitated in the early phase of concentration, preventing any abundance of alkaline earth chlorides later. Even gypsum deposition may not be significant if calcium is scarce and becomes eliminated in the precipitation of calcium carbonate. In a similar fashion, should there be sufficient calcium to precipitate all the sulfate, no sodium sulfate will be deposited during subsequent concentration. In this way the initial proportions of ions play a major part in determining the sequence of salt deposition during evaporation, and the theoretical sequence will be observed seldom, if at all. It should also be remembered that the solubility of a given salt is not the same in the presence of other salts as in distilled water (*cf.* Hem, 1959, p. 74, 78, 81, 103).

Hutchinson (1957, p. 565) considers two examples of the changes taking place during evaporation of normal river waters, where salt beds and saline springs presumably are not important sources of ions. In the Lahontan basin and adjoining regions of western North America the most dilute waters are strongly dominated by carbonates, but where evaporation is more advanced the proportion of carbonate declines, that of sulfate increases slightly, and the proportion of chloride increases markedly. Finally, the most concentrated waters are greatly enriched in sodium and chloride, which become the most abundant ions by far, whereas sulfate makes up only a very small proportion of total anions. Some lakes of Indian Tibet, whose initial stage is likewise one of strong carbonate dominance, develop marked sulfate dominance and carbonate impoverishment in the phases of greatest concentration. Hutchinson suggests that this is because calcium was scarce initially and little was left for sulfate precipitation following the stage of carbonate deposition.

Anomalous types of concentrated waters may arise where the geological situation is unusual and modifies climatic influences. For example, where rocks are particularly rich in alkalies and poor in alkaline earths, waters may be greatly enriched in sodium, bicarbonate, and carbonate ions, as are those of Moses and Soap lakes in the State of Washington, U.S.A. (Hutchinson, 1957, p. 568) and the streams of Mount Meru, Tanganyika (Sturdy, Calton, and Milne, 1933).

Closed lakes whose salts are derived from primary weathering of igneous rocks may differ chemically from those whose waters leach sedimentary strata rich in salts of various kinds, often of oceanic origin. Waters draining sedimentary rocks may exhibit a great variety of ionic proportions, as shown in the wide range of data compiled by Clarke (1924).

Yet another type of saline lake with salts of marine origin is exemplified by Sambhar Lake in northern India (Holland and Christie, 1909). Its ions apparently derive from continued aerial transport of sea salt inland from the Rann of Cutch by winds. Lake Corangamite in Australia appears to be similar (W. G. Anderson, 1945).

Concept of an Average Fresh Water

The average fresh water of the world has been set up recently as a standard whose ionic proportions would gradually be approached by all waters if sufficient time were allowed, through the operation of ion-exchange processes between waters and the soils and rocks through which they percolate (Rodhe, 1949; Hutchinson, 1957, p. 556). This doctrine is even more extreme than the monoclimax theories of plant ecology and soil science, which assert that in time all plant communities and soils should approximate norms determined solely by the climatic factor. Recent work suggests that such idealized situations are far from real, and that the complexity of both vegetation and soil reflects continual interaction of several partially independent ecosystem factors (Jenny, 1941; Major, 1951; Crocker, 1952; *see also* Billings, 1952). A similar interaction is surely responsible for the wide range of variation in the ionic composition of natural waters. Some examination of the lakes of eastern Sweden will show this to be true even of the area where the hypothesis of an ideal "standard composition" originated (Rodhe, 1949; *cf.* Eriksson, 1955, p. 249).

In the Swedish districts of Dalarna and Uppland, Lohammar (1938) analyzed a great many lake waters, and his data show that the average ionic proportions of 21 Uppland lakes closely resemble those of the average fresh water of the world (Rodhe, 1949). There is in these Uppland and Dalarna lakes an inverse relation between total salt concentration and

height above sea level (Gorham, 1955a, p. 142), attributed by the present author to greater leaching at higher altitudes. Eriksson (1955, p. 249) gives a much more likely explanation, that the relationship is due mainly to a marked concentration of salts by evaporation in the lowlands, where precipitation is much less than in the uplands. However, in discussing these lake waters further Eriksson adheres to Rodhe's hypothesis of uniform ionic proportions and lowest ratios (commonly less than 1) come from the very peculiar Dalarna seepage lakes on eskers. In these, calcium accounts for a much smaller, and potassium a much larger proportion of total cations than in other Swedish lake waters. Troedsson (1955, p. 179) records similar differences in cation balance between waters seeping into lysimeters (underground instruments for measuring percolation), or expressed from the soil by pressure, and the

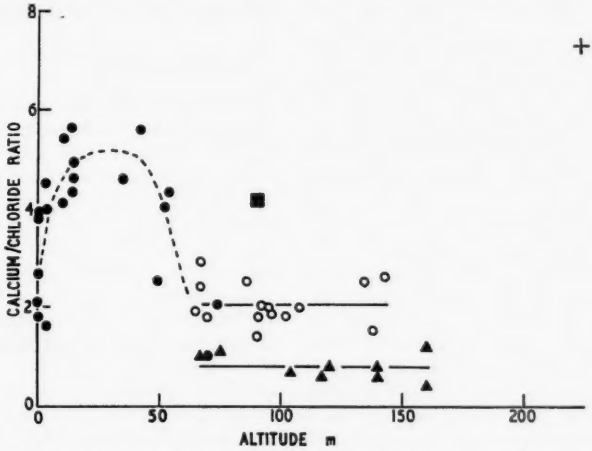

Figure 1. Ratio of calcium to chloride in Swedish lake waters of varying elevation. Solid circles—Uppland; open circles—Dalarna lakes on glacial till; solid triangles—Dalarna lakes on eskers; solid square—average for lakes on Dalälven sediments; cross—average for lakes on Siljan limestone

implies that the most significant factor controlling them is probably the composition of the precipitation, which remains unchanged throughout the upper parts of the soil where the ion-adsorbing colloids are in quasi-equilibrium with percolating rain water.

Such a view obscures the very real influence of soils and topography upon water chemistry, illustrated by the relation between calcium/chloride ratios and altitude of East Swedish lakes. In Figure 1 evaporation effects are excluded by calculating the change in calcium, the major cation in these waters and largely supplied from the soil, in proportion to the change in chloride, which is known to be supplied almost wholly from the atmosphere and to be little affected by biological uptake or soil fixation. As might be expected, the calcium/chloride ratio is highest (7.3) in lakes of the Siljan limestone area. In contrast, the common type of lake and ground water in this general area.

Among other Dalarna lakes there is a clear distinction between those on Dalälven river sediments and those on glacial till. The former exhibit calcium/chloride ratios averaging 4.2 and are both absolutely and proportionately richer in calcium, the source being presumably Siljan limestone drift brought down in late glacial times. The lakes on till show ratios of about 2.1, not greatly different from those of recent Swedish rain, and therefore suggesting that these lakes receive little calcium from soil leaching. In this group of lakes the constancy of the ratio at different altitudes favors Eriksson's hypothesis that an increase in the precipitation/evaporation ratio is chiefly responsible for the decline of total salt concentration with increasing elevation.

The Uppland lakes show most interesting

variation in calcium/chloride ratio. The waters highest above sea level have low ratios, similar to those of Dalarna waters on till at the same or greater elevations. At lower levels the ratio rises to about 5, presumably through solution of lime from calcareous clays and marls deposited during the marine submergence of this part of Sweden at the end of the ice age (Lundqvist, 1935, p. 293). Very near sea level distinctly lower ratios (in some cases less than 2) are again observed. These low ratios are due to a marked rise in the chloride concentrations of these waters, accompanied by a similar rise in sodium and magnesium. Marine influence is evident, but there seems little likelihood that differences in sea-spray supply are responsible, since neighboring lakes exhibit very different ratios. Because the lakes concerned are very near sea level, it seems reasonable to suppose that leaching has not yet eliminated the effects of marine submergence.

Such large variations in calcium/chloride ratio, even within the Uppland group of lakes, indicate the profound influence of topographical, geological, and historical factors upon the chemistry of waters in this region. Such effects can be obscured, however, by statistical treatment suggesting an ideal or "standard" composition resembling that of the average fresh water of the world.

BASIS OF PRESENT INTERPRETATION

If all fresh waters are not tending toward an ideal "standard composition," the factors operating to bring about and maintain differences in their ionic make-up must be examined. Five partially independent factors qualify for investigation: climate, geology, topography, biota, and time. It is evident that these factors are to some degree interdependent, and all may interact to condition the concentration and composition of any given body of water. Therefore, in order to demonstrate the effects of any factor, it is desirable to select (as far as possible) special cases in which that factor is of paramount importance and the others have little effect. Such instances are probably rare, and have been seldom sought, so that it will also be necessary to examine situations where factors interact to some extent. First, however, the sources and mechanisms of ion supply to natural waters may be considered, for knowledge of the processes through which environmental factors act may aid in the understanding of their effects.

SOURCES AND MECHANISMS OF ION SUPPLY

General Statement

It has been generally and quite correctly supposed that the main source of ions in fresh waters is the soil, weathered chiefly through attack by carbonic acid. However, it has become increasingly evident that a large part of the salts found in more dilute lake and stream waters, and even in some salt lakes of arid regions, may come from the atmosphere.

Illustrations of the Importance of Atmospheric Supply

In the English Lake District many small lakes on volcanic rocks contain water with less than 0.2 m equiv./l of total salts, to which rain probably contributes almost the whole of the chloride, sulfate, sodium, and potassium. More than 80 per cent of the magnesium and more than half the calcium in these lakes also appear to be supplied by rain; the remainder of these two ions is supplied as bicarbonates by soil weathering (Gorham, 1958a, p. 173). Even in the most concentrated waters of the area, which range up to as much as 5 m equiv./l and drain sedimentary rocks, almost all the chloride and sodium probably comes from the rain, together with roughly half the sulfate and about 20 per cent of potassium and magnesium. Calcium and bicarbonate, which strongly dominate the cations and anions of these richer waters, are supplied almost wholly by the soils and rocks, in many cases limestone.

In Finland too a good deal of the material carried by rivers probably comes from atmospheric precipitation. Using data from the quarterly appendices to the journal *Tellus*, the atmospheric supply of ions in rain and snow may be calculated over a full year 1956–1957 at Kuopio and Jyväskylä, and compared with the quantities removed (per unit area) by the river systems Vuoksi and Kymijoki (Viro, 1953, p. 27), within whose drainage basins the two precipitation stations lie (Table 4). Such a comparison, despite inadequate precipitation data, shows the significance of the atmosphere as a source of ions. In these rivers the air appears to supply about 20 per cent of the sodium, 10–20 per cent of the potassium, about 10 per cent of the magnesium, 20–40 per cent of the calcium, 30–50 per cent of the chloride, and 50–70 per cent of the sulfate. In reality almost all the chloride probably comes from the atmosphere, since the rocks

are very deficient in it, and the present method of calculation greatly underestimates not only chloride but also probably other ions from precipitation. The discrepancy may well be due to the fact that vegetation captures drifting salt and dust particles from the air during dry weather, which are subsequently washed off and into the soils and streams during rainy periods. Eriksson (1955, p. 245; 1960, p. 80) has suggested this explanation to account for a similar discrepancy between chloride discharge and precipitation in Sweden, while region; in some cases major-ion content in the rain is actually higher. Data for all these areas are included in Table 5.

Another indication of the importance of atmospheric ion supply, even to large continental river systems, is given by Alekin and Brazhnikova (1957), although they admit their data from the USSR give an exaggerated idea of its significance, owing to recycling of wind-blown salts from the soil during dry weather. (*Cf. also* Hutton and Leslie, 1958, p. 504.)

Concerning saline waters in arid regions, the

TABLE 4. SUPPLY OF IONS BY ATMOSPHERIC PRECIPITATION, AND THE DISCHARGE OF IONS BY RIVER WATERS, IN TWO DISTRICTS OF FINLAND

Precipitation data from Kuopio and Jyväskylä cover the period July 1956 to June 1957. In the former case March 1958 was substituted for March 1957, when some analyses were lacking; precipitation was very low in both these March samples.

	Ca	Mg	Na	K	SO_4	Cl
			kg/ha/yr			
Precipitation, Kuopio	2.7	0.4	1.0	0.6	9.3	1.5
River discharge, Vuoksi	12.2	3.9	6.0	5.4	17.7	4.9
Precipitation, Jyväskylä	4.2	0.4	1.4	0.9	11.0	3.2
River discharge, Kymijoki	10.6	3.6	5.6	4.1	15.5	6.4
			precipitation as % river discharge			
Kuopio/Vuoksi	22	10	17	11	53	31
Jyväskylä/Kymijoki	40	11	25	22	71	50

Tamm (1953, p. 88) and Madgwick and Ovington (1959, p. 17) provide data on rain composition, in the open and under trees, that support such a view. It would apply especially in dry but well-forested areas.

Data from Swedish lakes (Lohammar, 1938) and rain stations (quarterly appendices to *Tellus*) show that minimal concentrations of major ions in the lake waters are close to average levels in precipitation, even when evaporative concentration of the latter is disregarded. Analyses of Wisconsin lake waters (*see* Hutchinson, 1957, Table 67, p. 551) give a similar picture of minimal concentrations near those reported for precipitation from the same general region by Junge and Werby (1958). The last-named authors' analyses of precipitation in Newfoundland are also closely similar to the averages for analyses by Gorham (1957b, p. 14) of lake waters on granite in Nova Scotia. Harrison's (1934, p. 15) analyses of rainfall at Georgetown, British Guiana, are likewise not greatly different from analyses cited by Clarke (1924, p. 95) for river waters from the same data of Holland and Christie (1909) and V. G. Anderson (1945) indicate that aerial salt supply may sometimes be of great significance.

Eriksson (1960, p. 72) has attempted a balance sheet for supply of elements (S, Cl, Na, Mg, Ca, K) to the ocean, in which two estimates are given for air-borne material.

Types of Atmospheric Supply

Review of literature. Although atmospheric supply of ions to lake and river waters has seldom been considered of great importance, the mineral materials in the atmosphere that are precipitated from it have long engaged attention (*see* Eriksson, 1952), and a great deal of information has been gathered about them (reviewed by Junge, 1958). The concentrations of the major ions present in air and in precipitation are being intensively studied in Europe (*see* quarterly appendices to *Tellus*), and some work is being done in the United States. (*See* Junge, 1956; Junge and Werby, 1958.)

Rain. Much of the atmospheric salt reaching lake and river waters is washed down by

rain. In wet areas such as the English Lake District the amounts supplied to the ground daily are much more closely related to rainfall than to any other variable (Gorham, 1958a, p. 150), and, in Australia and Hawaii, Twomey (1955) and Squires and Twomey (1957) have demonstrated the influence of precipitation in removing salt particles from the air.

portance of dry fallout, this time from more arid continental areas, may be seen by comparing the concentrations of salts in precipitation falling upon inland areas in the central U.S.A. (Junge and Werby, 1958) and the U.S.S.R. (Alekin and Brazhnikova, 1957). Junge and Werby, using collectors open only during intervals of precipitation, report concen-

TABLE 5. COMPARISON OF AVERAGE IONIC CONCENTRATIONS IN 1 YEAR'S ATMOSPHERIC PRECIPITATION WITH AVERAGE OR MINIMUM CONCENTRATIONS OBSERVED IN LAKE OR RIVER WATERS (ppm)

		Ca	Mg	Na	K	SO_4	Cl
Precipitation	Single station, Newfoundland	0.8	n.d.	5.2	0.3	2.2	8.9
Lake water	Winter average, waters on granite, Nova Scotia	1.0	0.5	5.2	0.4	5.9	7.7
Precipitation	Single station, Wisconsin	1.2	n.d.	0.5	0.2	2.9	0.2
Precipitation	Single station, Minnesota	1.0	n.d.	0.2	0.2	1.4	0.1
Lake water	Minimum concentration observed, Wisconsin	0.1	<0.5	0.1	0.3	0.8	0.1
Precipitation	Single station, Öjebyn, northern Sweden	1.2	0.2	0.4	0.3	2.5	0.7
Lake water	Minimum concentration observed, northern Sweden	1.1	0.5	0.6	<0.2	0.2	0.4
Precipitation	Single station, Ultuna, central Sweden	0.6	0.1	0.3	0.2	2.6	0.5
Lake water	Minimum concentration observed, Dalarna	0.6	0.3	0.3	trace	0.2	1.3
Lake water	Minimum concentration observed, Uppland	1.7	0.6	1.0	<0.4	2.4	1.7
Precipitation	Three-year average, Georgetown, British Guiana	0.8	0.3	1.5	0.2	1.3	2.9
River water	Demerara River	0.3	1.9	8.0	1.2	0.8	7.5
River water	Essequibo River	1.2	0.9	2.1	0.2	1.7	1.0

Dry fallout. In less humid regions dry fallout may be appreciable; in Massachusetts, for example, Junge and Gustafson (1957, p. 164) found about 25 per cent more chloride in continuously open containers than in containers open only during precipitation periods. In Japan Hanya (1951) calculated that rainfall carries down only about 14 per cent of the chloride in river waters, while 41 per cent is due to dry fallout. As mentioned earlier, Eriksson (1955; 1960) has suggested that vegetation is an important collector of dry salt particles from the air of Sweden, where the amount of river-borne chloride greatly exceeds that supplied by precipitation and cannot be accounted for by rock weathering or other sources. Yet another indication of the im-

trations much below those recorded by the Soviet workers, who apparently left their collectors open continuously. Perhaps the most impressive examples of dry fallout are the salt lakes of the Sambhar region in India (Holland and Christie, 1909), where very high salinities appear explicable by aerial transport of sea salt during the dry season from the Rann of Cutch, more than 500 km away. Eriksson (1958, p. 169) has recently dealt with the general question of sea-salt transport into arid regions.

To illustrate the concentrations of ions found in air, Table 6 presents the range of values given in the 1958 issues of *Tellus* for 72 European stations during the month of December 1957 and 69 stations during the month

of June 1958. The air samples are very pure indeed at some stations, while large amounts of ions are present at others, especially near the sea or near industrial centers.

Snow. In high latitudes snow is an important fraction of precipitation, and this may lessen the amount of salt carried down from the air. Total ash, sulfur, and nitrogen are much lower in snow than in winter rain at Kentville, Nova Scotia, suggesting a lesser efficiency of snowflakes in removing materials from the air (Herman and Gorham, 1957). Denisov and Bugaiov (1956) report much smaller differences from the U.S.S.R., where, however, the precipitation is unusually rich in salts, probably owing to dustfall. Even here "dry" snow has a lower mineral content than "wet" snow.

Gases. All the major anions in natural waters may be cycled through the atmosphere, or at least supplied to it, partly in a gaseous state. Bicarbonate of course derives from carbon dioxide, and sulfur may be present in air either as sulfur dioxide or hydrogen sulfide. Under certain circumstances (Gorham, 1958a, p. 166, 1958d) chlorine too may be released to the atmosphere as a gas, hydrogen chloride, which soon dissolves in cloud and rain droplets. Junge (1956, p. 137) has claimed that the air may carry as much sulfur and chlorine in the gas phase as in solid or dissolved particulate form.

Table 6. Range of Ionic Concentration in European Air Masses During December 1957 and June 1958
In $\mu g/m^3$ or kg/km^3

	No. of stations	S	Cl	Na	K	Mg	Ca
Dec. 1957							
Min.	72	0	0.6	0	0	0	0
Max.		67	240	212	24.6	20	31
June 1958							
Min.	69	0	0	0.2	0.3	0	0
Max.		41	89	56	6.0	10	33

Sources of Atmospheric Supply

The sea. Chloride, sulfate, sodium, and magnesium are supplied to the air in large amounts from sea spray. The way in which sea salt enters the atmosphere is thought to be as follows, according to the views of Woodcock (1953, p. 364; *see also* Owens, 1940). High winds lead to wave formation on the sea surface, and the larger waves tend to break and produce foam patches containing a multitude of minute bubbles. Such foam patches may also be formed by waves breaking on the sea shore (Boyce, 1951). When the bubbles surface and burst, they shoot up tiny jets which break into droplets at their tips. Many of the droplets are sufficiently small to be swept up into the air, and may evaporate there to form salt particles, which are capable of acting as nuclei for cloud and raindrop condensation (Woodcock *et al.*, 1953), or of being captured by raindrops. Apparently about 99 per cent of the mass of natural sea spray is found in the so-called "giant particles" above 0.8 μ radius (Junge, 1957, p. 31), and these are large enough to be swept from the air in considerable amounts by vegetation (Sugawara, Oana, and Koyama, 1949; Boyce, 1954, p. 41; Eriksson, 1955, p. 245; 1958, p. 152). According to Sugawara (1951, p. 14) the chloride content of streams may vary under different types of vegetation cover because of variation in the effectiveness of the plant cover in filtering salt from the air.

The actual presence of sea-salt particles in the air, even far inland, has been shown frequently, for example by Woodcock (1953), Twomey (1954), and Junge (1954; 1956). Their connection with cloud and raindrop formation has been suggested by Woodcock (1952; Woodcock and Blanchard, 1955). Twomey (1959, p. 241), however, claims that the main source of cloud nuclei is probably the dry land surface; Went (1960) speculates that volatile plant products may also serve in this capacity, and Bowen (1956) suggests meteoritic dust for the same role.

Many studies showing a marked decline of

rain, river, and lake water chlorides inland point to sea spray as a very important source of this ion in maritime areas. Studies on the ratios of sodium and magnesium to chloride in rain from the English Lake District (Gorham, 1955b, p. 232), and on the influence of winds from the sea on rain chlorides (Gorham, 1958a, p. 156), provide further evidence on this question. Perhaps the purest example of marine salt precipitation is the Norwegian rime analyzed by Köhler (1923), for which data are given in Table 7. Rain at Cape Bridgewater in South Australia (Hutton and Leslie, 1958, p. 500) likewise appears to contain rather pure sea water, except for the addition of considerable calcium and bicarbonate from calcareous sand dunes (Hutton, 1958, p. 289). Analyses of this rain are also given in Table 7. Even far at sea, however, particulate matter in the atmosphere may be very different from sea salt in ionic composition. Lodge, Macdonald, and Vihman (1960) report that sulfate exceeds chloride by weight in air samples taken about midway between San Francisco and Honolulu.

Table 7. Composition of Rime at Hallde in Norway and Rain at Cape Bridgewater in South Australia, in Comparison with the Composition of Sea Water

Chloride taken as 100 in each case; calculations by weight

	Ca	Mg	Na	K	SO_4	Cl
Sea water	2.1	6.7	56	2.0	14	(100)
Hallde rime	2.4	7.6	56	n.d.	16	(100)
Cape Bridgewater rain	6.4	7.3	57	2.2	9.6*	(100)

* The sulfate methods employed for these Australian analyses were said to be generally unsatisfactory for low amounts. If sulfate is calculated by difference (sum of cations minus sum of chloride and bicarbonate), the value in this table would be 14, identical with that of sea water.

If the "giant particles" were the only source of sodium and chloride in atmospheric precipitation, the two ions should be present in the same proportions as in sea water. However, the chloride/sodium ratio frequently departs far from the sea-water value of 1.8, sometimes with chloride and sometimes with sodium in marked excess (Rossby and Egnér, 1955; Gorham, 1958a, p. 161). Soil dust probably contributes excess sodium (Kalle, 1953–1954; Junge and Werby, 1958, p. 423), which Hutton (1958) has shown to be correlated with excess calcium and potassium in South Australian rain. Volcanoes (Bottini, 1939) and particularly industrial areas (Gorham, 1958a, p. 161; 1958d) can supply excess chloride. Cauer (1949, p. 229; 1951, p. 1131) has suggested that sea-salt particles may be oxidized by atmospheric ozone, releasing free chlorine gas which is eventually transformed to hydrochloric acid, and substituting bicarbonate for chloride in combination with the sodium ions. This process is not likely to be quantitatively important (Eriksson, 1960, p. 77), although it has been argued recently by Oddie (1959). A greater likelihood of decomposition would appear to arise from the mixing of sea-salt particles with sulfuric-acid droplets from centers of air pollution or other sources. Upon abstraction of moisture by freezing or evaporation, gaseous hydrochloric acid could be released from the mixed droplets, thus altering the balance of chloride and sodium in precipitation. The droplets would exhibit an excess of sodium, but the hydrochloric acid would eventually enrich precipitation elsewhere in chloride. Some experimental and observational evidence for such a process has been adduced by Gorham (1958a, p. 165), and it has also been suggested independently on theoretical grounds by Eriksson (1958, p. 167; 1960, p. 77).

Erikksson (1958, p. 154) has further suggested that a similar process of chloride release may take place through the absorption and subsequent oxidation of sulfur dioxide by arid soils rich in sodium chloride. However, arid soils are commonly quite alkaline, and would probably neutralize any acid as soon as produced. Such release would be more likely from strongly acid forest soils, which may be rich in sulfuric acid (Gorham, 1958b; 1958f) and often become very dry at the surface in summer.

According to Robbins, Cadle, and Eckhardt (1959) the release of hydrogen chloride from

sodium chloride particles might also be accomplished through interaction with nitric acid produced by hydrolysis of nitrogen dioxide.

Sugawara, Oana, and Koyama (1949, *see also* Koyama and Sugawara, 1953) have suggested that evaporation of sea salt in the air may lead to a separation of hygroscopic chloride particles, easily washed out by rain, from less hygroscopic and consequently more long-lived sulfate and carbonate particles. This hypothesis deserves more investigation, particularly by experiment. It has been discussed recently by Eriksson (1960, p. 76), who doubts its importance.

Conway (1942, p. 156) has postulated the sea as a source of more sulfate than that contributed by sea spray. To account for a large deficiency in the balance sheet for sulfur in river waters, he invoked diffusion of volatile sulfur compounds (presumably hydrogen sulfide) from the muds of the continental shelf into the sea, and thence to the atmosphere, where they are oxidized to sulfate and washed down by rain. Such sulfate might provide condensation nuclei for the sea mist causing atmospheric opacity at Valentia in western Ireland. Wright (1940) found that the mist nuclei there are definitely hygroscopic; he believed them to be particles of sea salt. However, Simpson (1941b) showed that the nuclei could not be sea salt, for almost no reduction in air opacity occurred as relative humidity dropped below 70 per cent. Because this is about the point at which sea salt nuclei crystallize, a sudden reduction in opacity should occur if sea-salt droplets are responsible. Simpson (1941a, p. 120) concluded that nitrous acid might be involved, since industrial sources of sulfate were not present in the area, but if Conway's hypothesis is correct the sea floor could be an alternative source of sulfate. Yet another nonindustrial source might be the extensive water-logged peat deposits of western Ireland, from which some release of gaseous hydrogen sulfide is most likely.

The possibility that hydrogen sulfide is released from the sea floor has been discussed recently by Eriksson (1960, p. 74), who suggested that very small concentrations and gradients, undetectable by present analytical techniques, could produce enough sulfate to balance the sulfur not otherwise accounted for in river waters. He also remarked that tidal flats may be a source of hydrogen sulfide, and that it could be released in large amounts by protein breakdown in the metabolic turnover of organic matter near the sea surface.

The recent data of Junge and Werby (1958, p. 422) have been used to both deny and assert the importance of the sea as a source of hydrogen sulfide, which eventually oxidizes to sulfate in the atmosphere. Junge and Werby (1958, p. 423) presented a map of the concentration of excess sulfate in precipitation over the United States, for which the values were calculated by subtracting from observed sulfate the amount expected from the observed chloride concentrations if the two ions were present in the same proportions as in sea water.

Table 8. Contents of Organic Matter (Ignition Loss), Ash (Ignition Residue), Silica, Calcium, and Potassium in 10 Samples of Finnish Snow (ppm) (Viro, 1953)

Ignition loss	Ignition residue	SiO_2	Ca	K
1.58	1.24	0.14	trace	0.10
2.36	1.26	0.23	0.03	0.09
3.89	2.48	0.27	0.07	0.19
3.28	2.14	0.36	0.13	0.11
4.36	5.88	0.42	0.25	0.63
5.85	6.60	0.45	0.34	0.83
8.17	10.02	0.52	0.79	0.51
5.90	9.15	0.62	0.30	0.74
4.00	4.21	0.66	0.41	0.20
8.70	12.73	1.44	1.14	0.93

This map showed greater excess sulfate concentrations inland, which was taken to indicate that the source is the land surface. Eriksson (1960, p. 91) used the same data to calculate the excess sulfate supplied per unit area of land surface and indicated a tendency for the amounts to be higher near the sea. The suggested coastal gradients are, however, not very clear. Since both ion concentration in precipitation and ion supply to the ground are strongly influenced by the amount of precipitation and other environmental factors, it is difficult to place much reliance on either of the above interpretations.

Soil. In addition to being recycled in the dry state through the atmosphere, dust from the ground makes an important contribution of salts to rain and snow in many cases. It seems to be a major source of calcium and potassium in precipitation, as has been emphasized by Kalle (1953–1954), Junge and Werby (1958, p. 421–423), Durov and Fedorova (*see* Chilingar, 1956), Hutton (1958), and Hutton and Leslie (1958, p. 504). Analyses of snow by Viro (1953, p. 36) exhibit distinct correlations between calcium, potassium, and the silica con-

tent of the snow (Table 8), which might be taken to suggest the importance of soil dust. Since there is also a marked correlation with organic matter, such an interpretation remains doubtful. The presence of appreciable amounts of calcareous dust in the air near limestone regions has also been inferred from pH values and calcium concentrations in the rain of Sweden (Barrett and Brodin, 1955, p. 256) and western Ireland (Gorham, 1957a).

Table 9. Percentage Composition of Dust from Greenland Ice

From Tamm (1958)

SiO_2	62.25
Al_2O_3	14.93
Fe_2O_3	0.74
FeO	4.64
MnO	0.07
CaO	5.09
MgO	3.00
Na_2O	4.01
K_2O	2.02
P_2O_5	0.11
Cl	0.06
H_2O	0.34
Ignition loss	2.86

Rankama and Sahama (1950, p. 310), Eriksson (1952, p. 296), and Tamm (1958, p. 238) have reviewed the occurrence of occasional heavy dustfalls over large areas, and Tamm (1958, p. 235) quotes an analysis by Free of dust collected from Greenland ice, which may be considered representative of long-distance fallout (Table 9). According to Tamm and Troedsson (1955, p. 67) the larger part of the potassium in dust, as well as appreciable fractions of the calcium, may not be readily soluble in water, but almost all the sodium is water soluble.

Evidence concerning the actual presence of soil dust in normal air masses has been provided by Twomey (1953; 1954), who found it most plentiful during anticyclonic summer weather in Australia, and more abundant in offshore than onshore air streams. Junge (1954, p. 333) also observed insoluble particles in atmospheric aerosols from Massachusetts, even in polar air masses almost devoid of soluble salts. Some reddish-yellow dust particles observed in a few samples from Florida (Junge, 1956, p. 129) were believed to have originated in west Africa and to have travelled across the Atlantic beneath the inversion layer in the atmosphere.

Volcanoes. Over the ages volcanoes have been major sources of the volatile elements carbon, chlorine, and sulfur, which emanate in the form of carbon dioxide and monoxide, hydrochloric acid, hydrogen sulfide, and sulfur dioxide and trioxide (Rankama and Sahama, 1950, p. 184; Rubey, 1951). Local influence of volcanic activity upon rain chemistry is illustrated by analyses of rain water from near Vesuvius (Bottini, 1939), which show a very close correlation between acidity and chloride concentration, the maximum level of hydrochloric acid being 1.8 m equiv./l.

Air pollution. Another source of atmospheric ions, very important at present, is industrial and domestic air pollution. Meetham (1950, p. 362) compares deposits of total ash, combustible material, sulfates, and chlorides in British urban and rural areas; and the recently established European network for air and precipitation analyses (*see* quarterly appendices to *Tellus*) is providing many more data on air-pollution effects, although unfortunately no analyses are being made within cities. Many deposit gauges are, however, located in British cities, and their contents are analyzed for total ash and ignitable matter, sulfate, and chloride, and in many cases for calcium and pH as well (Dept. of Scientific and Industrial Research, 1955). In the U.S.A. a program of urban air analysis is now well established (Anonymous, 1958).

Among industrial sources of atmospheric pollution, metal smelters rank as heavy emitters of sulfur dioxide (Katz, 1939). In the Sudbury area of Ontario, where roughly a million tons of sulfur escape to the air annually, fallout is mainly restricted to within about 5 miles of the three smelters and is quantitatively negligible beyond about 15 miles (Gorham and Gordon, 1960).

Junge (1954; 1956) has examined the nature of aerosol pollutants and finds the particles usually of less than 0.8 μ radius. Ammonium sulfate is a major component, and ammonia may be important in facilitating oxidation of sulfur dioxide to sulfate (Junge and Ryan, 1958). Herman and Gorham (1957) reported a marked correlation between ammonia and sulfate in acid rains and snowfalls from Nova Scotia, which parallels Junge's results with the aerosol particles. Other analyses of rainfall (Houghton, 1955; Gorham, 1955b, p. 236; 1958a, p. 160) indicate that free sulfuric acid is also of great importance. Twomey (1953) has observed droplets in the air near Sydney,

Australia, which remained liquid at 20 per cent relative humidity and appeared to consist of sulfuric acid.

Oxidation of sulfur dioxide to sulfate may take a considerable time, and sulfur may diffuse in the gaseous phase far from its industrial sources (Junge, 1956; Junge and Ryan, 1958; Meetham, 1950, p. 369; 1954). Moreover, sulfides are present in soot and can probably oxidize slowly to sulfate. Material from a black soot film, formed on the surface of Windermere in the English Lake District during calm weather, gave a noticeable odor of hydrogen sulfide upon acidification and a distinct reaction with lead acetate paper (Gorham, unpublished). Soot from a fireplace in the Windermere laboratory of the Freshwater Biological Association also contained sulfides.

Gorham (1958a, p. 161; 1958d) has suggested that hydrochloric acid may be a frequent component of air pollution, providing excess chloride to raise the chloride/sodium ratio of rain above that of sea water (Gorham, 1958a, p. 163; Junge and Werby, 1958, p. 419). Such a source probably accounts for the unusually high chloride/sodium ratios observed by Rossby and Egnér (1955) in precipitation from warm and moist air masses entering Sweden from the south and southeast.

In addition to sulfuric and hydrochloric acids, air pollution may add appreciable amounts of calcium and potassium to rain (Gorham, 1955b, p. 236; 1958a, p. 160). Analyses by MacIntire and Young (1923) of rain in and just outside Knoxville, Tennessee, show considerable local additions of sodium and magnesium as well (Table 10). These authors also analyzed a sample of soot, which was exceptionally rich in calcium and slightly alkaline. Gorham (1958a, p. 162) analyzed materials dissolved from samples of chimney soot in London, England, which are especially rich in calcium and sulfate, but contain appreciable amounts of the other major ions as well (Table 11). These soot extracts are slightly acid and have a distinct excess of anions over cations. Presumably the balance is made up largely of ammonium, although trace elements (*e.g.*, antimony, arsenic, cobalt, lead, manganese, nickel, zinc) may be abundant in the particulate matter of urban air in England (Stocks, 1958). Lead and zinc are exceptionally plentiful, the former averaging 15–59 and the latter 5–33 $\mu g/100\ m^3$ in 1 year's samples from 10 towns and cities. Bear (1954) provides analyses of trace elements (copper, cobalt, manganese, molybdenum, zinc) in rain falling at New Brunswick, New Jersey, close to industrial centers, and numerous trace-element analyses are available for urban air in the U.S.A. (Anonymous, 1958).

Table 10. Ionic Composition of Rain at Knoxville, Tennessee (ppm)
(MacIntire and Young, 1923, as recalculated by Conway, 1942, p. 145)

	Ca	Mg	Na	K	SO_4	Cl
City	2.8	0.7	3.6	5.5	14.5	6.5
Outside city (2.4 km from center)	0.9	0.4	1.5	2.6	5.6	3.3

Organic matter. A last source of soluble ions in atmospheric precipitation is organic debris floating in the air. That this may be quite significant is shown, for instance, by Tamm (1958, p. 237) and Viro (1953, p. 36), who report ignition losses for rain and snow solids comparable to the ignition residues. Unpublished data for the Nova Scotian precipitation samples (18 per cent snow) analyzed by Herman and Gorham (1957) showed average ignition residues of 10 ppm, and losses of 13 ppm; 17 separate samples of snow averaged 2.7

Table 11. Soluble Materials in Five Samples of Smoke Solids from London Chimneys
(Gorham, 1958a)
Extract pH 5.6

	mg/g dry wt
Cations	
Ca	27.2
Mg	0.9
Na	7.4
K	3.9
Anions	
SO_4	81.6
Cl	27.0
NO_3–N	0.4

ppm ignition residue and 4.3 ppm ignition loss (the samples were not weighted by volume for averaging). Unfortunately there has been no attempt to correlate ionic composition and concentration with the levels of organic matter in precipitation, except for the rather crude correlation established between soot and sulfate in rain from the English Lake District (Gorham, 1955b, p. 234). A close relationship exists between ignition loss and ignition residue in the Finnish snow samples of Viro (1953, p. 36) (Table 7), so that independent correlations with ionic concentrations are difficult to establish in this instance. This was not the case with the Nova Scotian samples where, however, detailed ion analyses were lacking.

Neumann, Fonselius, and Wahlman (1959, p. 138) have reported concentrations of organic carbon in atmospheric precipitation in Sweden ranging from 0.8 to 3.4 ppm.

Ion Supply by Soil and Rock Weathering

General discussion. The capacity of rocks and soils to supply ions by weathering has been investigated in great detail, not only with regard to the nature of the material but also to the influence of various environmental factors upon the weathering process. However, it has not so far been possible to follow separate rainfalls through the soil phase of the hydrologic cycle, in order to evaluate quantitatively the relative influence of various weathering factors, sorption, plant uptake, etc., upon their ionic composition. Therefore only a brief account of weathering and related phenomena is given here, with special attention to aspects where the connection with water chemistry is clear. It is convenient to treat the weathering process under four headings—solution, oxidation or reduction, the action of hydrogen ions, and complex formation. A useful text for a detailed discussion is that of Keller (1957).

Solution. Ordinary solution not involving hydrolysis or acid weathering is important chiefly in sedimentary deposits rich in soluble salts. In the leaching of marine salt beds solubility differences may lead to relative enrichment of alkali chlorides in the resultant lake waters (Clarke, 1924, p. 176). Such solubility differences may also be responsible for marked fluctuations in ionic proportions of river waters during periods of high and low water. During a flood, sodium chloride may be dissolved more rapidly than calcium sulfate from the soil, with the result that chloride may predominate over sulfate in flood waters even though it is less important under normal conditions (Clarke, 1924, p. 72).

Some anomalous chloride/sulfate ratios shown by V. G. Anderson's (1945) data for South Australian stream waters may be explicable on a similar basis. When this ratio is plotted against chloride content (Fig. 2), a distinct peak occurs in the vicinity of 400 ppm of chloride. It seems possible that this level of chloride is reached by streams fed chiefly from rapid surface runoff over arid saline soils, with the very soluble chlorides being dissolved much more readily than the less soluble sulfates. The higher levels of salinity above 500 ppm of chloride could, on the other hand, represent streams percolating more slowly through the soil, and hence able to pick up relatively greater amounts of the slowly soluble sulfates. Even in such waters the chloride/sulfate ratio is somewhat higher than in the sea, which Anderson claims quite reasonably as the source (via air-borne spray) of the salt. The very dilute stream waters of the Murray River area appear greatly enriched in sulfate relative to chloride (average chloride and sulfate concentrations 1.9 and 0.9 ppm, chloride/sulfate ratio 2.2 as compared with 7.2 in sea water). Inasmuch as the coastal samples of rain from Victoria analyzed by Hutton and Leslie (1958) are if anything impoverished in sulfate, this enrichment (assuming the analyses are reliable at such low concentrations) presumably comes from local sources in the soil.

A different situation is described by Durum (1953) for the Saline River in Kansas. Surface runoff predominates during floods, and the dissolved salts of the relatively dilute river water are dominated by calcium and bicarbonate; during periods of lower flow, however, groundwater influence becomes stronger, and sodium, chloride, and sulfate make up the greater part of the more concentrated salt solution. When flow is minimal the salt concentration in the river water is maximal, and chloride exceeds sulfate in concentration.

Solubility differences may also be responsible for the relative paucity of magnesium and potassium in the Sambhar Lake of northern India, as suggested by Holland and Christie (1909, p. 167). This lake is said to owe its high salinity to wind-borne sea salt evaporated during the dry season on the coastal flats of the Rann of Cutch. As sea water evaporates and sodium chloride is precipitated, the remaining bitterns become relatively enriched in magnesium and potassium. If they, being strongly

deliquescent, remain largely unevaporated, then the salt particles blowing from the Rann of Cutch to the Sambhar region will be relatively deficient in magnesium and potassium, as in fact the lake water is.

Eriksson (1958, p. 172), on the contrary, claims that aerial supply of sea salt to Sambhar Lake may be adequate without invoking salt entrainment by winds sweeping over the dry coastal flats. He further remarks that carbonate precipitation, certainly responsible to some extent for the low calcium concentration in the Sambhar Lake water, may also lower the magnesium level. However, it would be unlikely to affect potassium, and the very low concentration of this ion would appear to favor the hypothesis of Holland and Christie that a good deal of the salt comes from the coastal flats during the dry season. Very small condensation nuclei can apparently be released in large numbers by fragmentation during the crystallization of large salt droplets (Twomey and McMaster, 1955), which might aid considerably in the escape of salt from the dry flats.

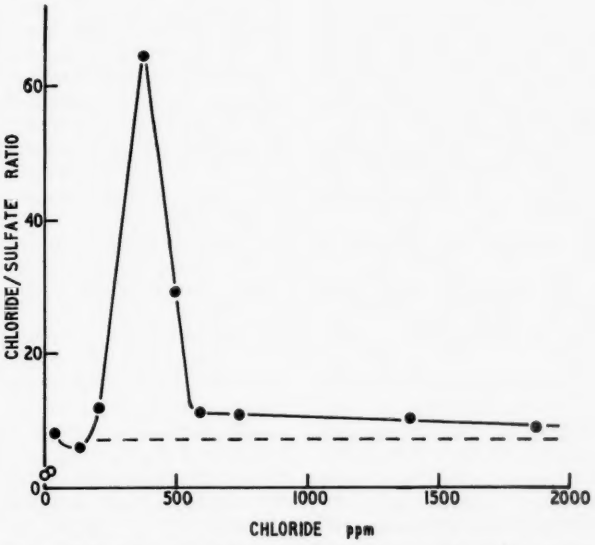

Figure 2. Weight ratio of chloride to sulfate in some Australian stream and river waters. Open circles mark the Murray River basin. Broken line—ratio in sea water

Oxidation and reduction. These processes affect chiefly iron, manganese, sulfur, nitrogen, phosphorus, and carbon compounds in soils. Sulfides of iron are common constituents of rocks and water-logged soils, and oxidation of these sulfides can be an important source of sulfate to natural waters, commonly in the form of sulfuric acid whose strong solvent properties lead to the supply of other rock and soil constituents.

Ferrous sulfide is frequently abundant in lake muds, in the surface layers of which a recurrent cycle of oxidation and reduction may take place during the phases of isothermal circulation in winter and of stratified stagnation in summer (Mortimer, 1941–1942; *see also* Gorham, 1958e). Wiklander and his associates (1949; 1950) state that relict lake soils (gyttja) in Sweden may be very rich in ferrous sulfide, especially the saline types. They have also studied the oxidation of sulfide in such soils and find the primary stage of chemical oxidation to free sulfur rapid, while the secondary oxidation of free sulfur to sulfate appears mainly microbiological and comparatively slow.

Oxidation of ferrous iron present in silicates may also bring about ion supply to natural waters, by decreasing the stability of the

minerals and making them more susceptible to weathering. Reduction of ferric iron present in or coated upon some soil minerals (Carroll, 1958) might perhaps have a similar effect.

Microbiological breakdown of organic sulfur compounds (*see* Frederick, Starkey, and Segal, 1957) may also add sulfate to natural waters, again commonly in the form of sulfuric acid. This strong acid is responsible for much of the acidity of highly organic woodland soils (Gorham, 1958b; 1958f), of bog waters (Gorham, 1956b, p. 378; 1958b; Gorham and Cragg, 1960, p. 179) and also of peaty lake waters, where a distinct correlation between acidity, sulfate, and dissolved organic carbon has been described (Gorham, 1957b, p. 17).

Reduction may decrease sulfate concentration in natural waters, with a concomitant rise in alkalinity. It is particularly noticeable in the stagnant bottom waters of lakes during summer stratification, when bacteria have used up the free oxygen originally present and very low or negative redox potentials prevail at the mud–water interface (Mortimer, 1941–1942; *see also* Hutchinson, 1957, Chapter 13, p. 753).

Action of hydrogen ions. If free acids are absent, hydrogen ions for rock and soil weathering are provided by the dissociation of water molecules, in the process of hydration and hydrolysis (Robinson, 1949, p. 47; Keller, 1957, p. 21, 24). This process is believed to be especially significant for weathering in the tropics, where the rather dilute stream waters carry large proportions of dissolved silica, and the soils are usually lateritic and rich in iron and aluminium sesquioxides (Jenny, 1950, p. 59). Strongly acid podzols may occur on extremely base-deficient substrata in tropical lowlands, however, although they are more characteristic of cool temperate regions. Colorless "white-water streams" are apparently typical of lateritic regions, and "black-water streams" colored brown by dispersed humus compounds drain the tropical podzols (Richards, 1941). According to Sioli (1956) both have very dilute waters, the latter more strongly acid and deficient in calcium. (*See*, however, the Essequibo River in Table 3; this water is strongly dominated by bicarbonates and thus cannot be very acid).

Berg (In press) states that black waters are also characteristic of the streams of the central basin of the Congo, covered mainly by swamp forests. The variable amount of humic substances, originating from incomplete decomposition of vegetable material under swamp-forest conditions, is responsible for the low pH of these waters (3.5–5.2), whereas the poor solubility and small variety of soil minerals in that part of the world is responsible for the low mineral content of the Congolese stream waters (specific conductivity due to ions other than hydrogen ions averages about 10 micromhos, total hardness is equivalent to about 0.6 ppm calcium).

Carbonic acid is undoubtedly a major source of hydrogen ions for weathering. Carbon dioxide exists in normal air at a concentration of 0.03 per cent by volume, but volcanoes may raise concentrations locally and have undoubtedly been of great significance over the ages in supplying the air with this gas. Biological respiration greatly increases carbon dioxide in soil air; for instance, Romell (1922) recorded a maximum of 2.8 per cent in forest soils with hardpan zones, and of 5–6 per cent in very wet soils under spruce. For rain water Clarke (1924, p. 54) quotes Bunsen's analyses of dissolved carbon dioxide, which range from 2.14 per cent by volume at 20°C to 2.92 per cent at 0°C. At the other extreme, the interstitial water of anaerobic lake muds may contain as much as 5–35 per cent (Koyama, 1953, p. 117). The importance of carbonic acid in weathering is well illustrated by the high proportion of bicarbonate ions in most river waters. Even those of 50 ppm salinity or less and draining only igneous rocks are predominantly bicarbonate waters (Conway, 1942, Tables 7, 8, p. 134–135). The significance of biological carbon dioxide production is illustrated by Metzger (1928), who found consistently higher bicarbonate concentrations in soil samples close to plant roots than in samples farther from them.

Strong acids may also be important. Sulfuric acid, either from air pollution or from organic decay, may be abundant in rainfall and in certain types of soils (Gorham, 1958b; 1958f). Hydrochloric acid is sometimes plentiful in urban rain (Gorham, 1958d), and volcanoes can supply large amounts locally (Bottini, 1939). Nitric acid may be significant too, since oxidation of ammonia to nitrate is common in soils. Gorham (1958a, p. 175) has shown that nitrate is the most abundant anion in the soil solutions of mull types of woodland humus layer in the English Lake District.

Colloidal acids, whether acid clays or the so-called humic acids, may also supply large amounts of hydrogen ions for soil weathering, as may plant roots, whose adsorptive surfaces

are strongly saturated by these ions (Keller and Frederickson, 1952; Keller, 1957, p. 30). Tentative evidence suggests that the more primitive among the higher plants are more effective in such weathering (Lewis and Eisenmenger, 1948). Animals may also influence the weathering process. (*See* Lutz and Chandler, 1946, p. 70).

According to Berg (In press) most of the hydrogen ions in blackwater streams of the Congo appear to come from dissociation of humic acids, although strong acids are presumably also present in waters below pH 4. Shapiro (1957) has recently extracted strongly acid humic compounds from organic matter dissolved in North American lake waters.

Complex formation. Ions can also be removed from rocks and soils by chelation, or other forms of complexing. The ion is joined to the chelator, usually a complex organic molecule, by more than one valence bond, and is prevented from combining with other ions, although it remains in solution. The significance of chelation in weathering is not clear as yet, but, since complex organic molecules are common in surface soils and on rock faces covered by plants, it may be expected to play a part. Not only organic molecules deriving from soil humus, but also excretions from plant roots or micro-organisms, may be active complexing agents. The subject is briefly discussed by Schatz *et al.* (1954) and by Keller (1957, p. 33).

Significance of Ion Exchange

Exchange of ions between soils and the soil solution, or between muds and the overlying waters in lakes, also affects ion circulation, although an equilibrium may often be attained owing to the finite adsorptive capacity of the soil colloids. Thus in certain circumstances the overall composition of water percolating the surface soil may be only little affected by such processes (Troedsson, 1952; *cf. also* Eriksson, 1955, p. 249), although short-term disturbances and fluctuations do occur. The very process of weathering commonly involves some degree of ion exchange, with hydrogen ions replacing metal ions attached to the soil minerals.

Some of the acidity of bog waters may be due to exchange of metal ions in rainfall for hydrogen ions generated during organic decay and adsorbed on peat colloids, the latter being displaced into solution and lowering pH (Gorham and Cragg, 1960, p. 179). The predominance of hydrogen ions on the adsorption complex of very acid peats has been pointed out by Malmer and Sjörs (1955, p. 69) and by Gorham (1956a, p. 148). During dry weather another type of ion exchange may occur, in which oxidation of sulfide produces free sulfuric acid whose hydrogen ions then replace metal ions adsorbed by the peat, and so alter ionic ratios in the bog waters (Gorham, 1956b, p. 379). For example, during very wet weather the sodium/chloride and magnesium/chloride ratios in bog waters at Moor House in the English Pennines are very close to values for sea water (sodium/chloride in sea water 0.56, in bog water 0.53; magnesium/chloride in sea water 0.067, in bog water 0.063). In very dry weather, however, ion exchange greatly increases these ratios in bog water (sodium/chloride 0.73; magnesium/chloride 0.136). The magnesium/chloride ratio becomes especially high, because the divalent magnesium ion is much more strongly adsorbed by the peat colloids than is the monovalent sodium ion. If the metal ions adsorbed on peat colloids could all be released into the surrounding bog waters they would be greatly enriched (Malmer and Sjörs, 1955, p. 70; Gorham, 1956a, p. 148). Malmer and Sjörs find ratios of exchangeable to dissolved ions as follows: sodium 5–6, potassium 25–150, magnesium 90–240, and calcium 150–300. This accords with the expected order of adsorption.

In lakes the importance of ion exchange between muds and waters has been stressed by Mortimer (1941–1942), who suggested that during summer stagnation adsorbed bases may be released into anaerobic bottom waters by reduction and breakdown of an iron-silica-humus adsorption complex. Once aerobic conditions are restored at the onset of isothermal circulation in autumn, this complex will be reconstituted, and bases will be adsorbed from the water again. However, the evidence for this is circumstantial. The marked fall in alkalinity of bottom waters at the autumnal overturn, cited by Mortimer as evidence of base adsorption, may well be due mainly to dilution, because of the mixture of a very small volume of concentrated hypolimnion water with a very large volume of more dilute epilimnion water. In aerating anaerobic mud-water cores from the bottom of Esthwaite Water (the lake chiefly investigated by Mortimer), the writer could detect no significant change in the concentration of calcium, magnesium, sodium, and potassium by the time precipitation of ferric hydroxide was approxi-

mately complete (48 hours, as indicated by cessation of the decline in conductivity caused by oxidation of dissolved ferrous iron).

Nevertheless, a continual release of ions from mud particles to the water surrounding them probably does take place, for the interstitial water of the mud is commonly richer in salts than is the overlying water. Table 12 shows this clearly; the data for Esthwaite Water are new, and those for the Grosser Plöner See have been taken from Ohle (1955b). Hypolimnetic water just above the mud is also usually richer in salts than is surface water, as shown in Table 13, which compares cationic concentrations and proportions in the epi- and hypolimnion of Esthwaite Water with those of the interstitial water of the muds, and of the cations adsorbed on the mud colloids. This comparison, from previously unpublished data, shows that the relative importance of sodium, supplied almost wholly by atmospheric precipitation, declines from the surface waters to the cations adsorbed on the muds, while the importance of calcium, contributed mainly by soil weathering, becomes steadily greater. The enrichment of calcium in the mud water and in the hypolimnetic water strongly suggests displacement of calcium adsorbed on mud colloids into the surrounding water, followed by diffusion into the overlying hypolimnion. Such upward diffusion would be especially favored during stormy periods when turbulence stirs up the surface mud layers (Gorham, 1958e).

Table 12. Ionic Concentrations in the Bottom Waters of Lakes and in the Interstitial Waters of Their Muds

	Sum of four cations m equiv./l	Specific conductance, micromhos at 18°C	Ca	Mg	Na	K	HCO_3	SO_4	Cl
			ppm						
Esthwaite Water, England									
Bottom water	0.88	n.d.	11	1.4	4.7	1.1	n.d.	n.d.	n.d.
Mud water	1.21	n.d.	16	1.8	5.2	1.8	n.d.	n.d.	n.d.
Grosser Plöner See, Germany									
Bottom water	n.d.	387	49	5	n.d.	n.d.	165	28	36
Mud water	n.d.	460	65	8	n.d.	n.d.	205	46	56

The driving force for such a diffusion is probably the replacement of adsorbed calcium ions by hydrogen ions generated in the mud through oxidation of carbon or sulfur compounds. That carbonic acid is likely to be an important source of such hydrogen ions is suggested by recent studies of Mortimer and Mackereth (1958, p. 931) on the north Swedish lake Torneträsk, but oxidation of sulfide (either organic or inorganic) may well provide another. Gentle aeration of undisturbed mud-water cores from Esthwaite Water, without agitation of the mud surface, leads to considerable acidification, and the lowering of pH in water overlying the mud is closely correlated with a rise in sulfate concentration there. Starting with mud-water cores in which the water (after aeration) had a pH of 7.6 and a sulfate concentration of 0.2 m equiv./l, the minimum pH observed in four cores aerated for 135 days was 4.6, in a water with 1.3 m equiv./l of sulfate. In this unpublished experiment the logarithmic pH values were linearly related to the arithmetic values for sulfate in the core waters.

In contrast, Ohle (1955b) has carried out experiments from which he concludes that the calcium and bicarbonate concentrations of some soft German waters may have been reduced through ion exchange, adsorbed hydrogen ions on the rather peaty mud being replaced by calcium ions from the water. Such an exchange either with lake muds or with surrounding soils may be the cause of very low calcium concentrations (well below levels in average precipitation at several inland American stations) observed in some Wisconsin seepage lakes (Table 5).

In other experiments Koyama and Sugawara (Sugawara, 1951, p. 8) reported that the colloidal floc in lake waters can adsorb not only alkaline earths but also sulfates. After sedimentation these ions may be released into the

interstitial water of the mud by some sort of "ripening" process. Hesse (1958, p. 178) has suggested that diatomite in lakes adsorbs appreciable sulfate, and Carritt and Goodgal (1954) have shown sulfate sorption by Roanoke River solids; the reaction, however, is readily reversible with sulfate-free solutions.

Diagenesis of sediments may well result in ion exchanges, although little appears to be known of its significance in lakes. Powers (1957) discussed an interesting transformation of illitic to chloritic material in terrestrial clays transferred to an estuarine environment; magnesium was absorbed by the clays preferentially to potassium. According to Mason (1952, p. 136) diagenetic alteration of montmorillonite to illite may be the process mainly responsible for removal of potassium from natural waters; the change involves incorporation of this ion into the clay mineral.

The influence of ion exchange is less clear in lake waters than in deep-well waters, which percolate through long columns of soil and rock minerals with widely varying amounts and proportions of adsorbed ions. Softening of originally hard waters is of frequent occurrence (Renick, 1924; Riffenburg, 1925; Foster, 1950), and similar softening has been postulated for ground water supplying a very shallow well in sand dunes at Blakeney Point in Norfolk, England. At this site calcium is probably supplied almost entirely from shell fragments in the sand, and sodium and magnesium are supplied from sea spray; yet in the well water less than 1 m below the dune surface the calcium/bicarbonate ratio (in equivalents) is considerably less than unity, while sodium/chloride and magnesium/chloride ratios are much greater than in sea water (Gorham, 1958g, p. 378). The ion exchange involved is probably similar to that postulated for arid alkali soils of the *solonetz* type, common in dry continental climates. (*Cf.* Robinson, 1949, p. 387.)

TABLE 13. CONCENTRATION AND COMPOSITION OF CATIONS IN THE WATER AND ADSORBED ON THE MUD OF ESTHWAITE WATER

	Sum of four cations m equiv./l	Ca	Mg	Na	K
		m equiv.% of total cations			
Epilimnion water	0.77	58	14	26	3
Hypolimnion water just above mud	0.88	61	13	23	3
Interstitial water of mud	1.21	65	12	19	4
Adsorbed ions on mud colloids*	m equiv./100 g dry mud				
0 – 1.5 cm depth	20.8	75	12	10	4
1.5 – 3.5 cm depth	13.9	81	10	6	3
3.5 – 5.5 cm depth	14.7	81	11	6	2
5.5 – 7.5 cm depth	13.2	82	11	5	2
7.5 – 9.5 cm depth	14.1	80	13	5	2

*Exchangeable with ammonium acetate

Where fresh-water aquifers become contaminated by sea water, the contact zone may exhibit much higher concentrations of alkaline earths than would a simple mixture of the two waters, owing to displacement of adsorbed calcium and magnesium by sodium from the advancing sea water (Piper, Garrett, and others, 1953, p. 87, 105).

Mode of Water Percolation

Whether water flows as superficial runoff or as deep ground water may exert a profound effect upon ionic concentrations in river and lake waters. Durum's (1953) study of the Saline River in Kansas illustrates this point. He found that during 1947 and 1948 surface runoff accounted for 65 per cent of total stream discharge, but only 24 per cent of dissolved solids carried, since the ground water was much richer in dissolved materials. In Wisconsin, seepage lakes fed largely by runoff are extremely dilute; the mode for specific conductance (in general

proportional to total dissolved salts, *see* Rodhe, 1949) is 12 micromhos (Juday and Birge, 1933). Wisconsin drainage lakes, with inlet and outlet streams draining deeper soil horizons, are much more concentrated, with a specific conductance mode of 67 micromhos. Hooper (1956, p. 123) reports similar results for the alkalinity of lake waters in Michigan.

In contrast Pennak (1958) records considerably higher ash and organic-matter contents in Colorado montane (2500–3200 m) lakes with only temporary outlets than in ordinary drainage lakes there. Probably the latter, because they are deeper, are less subject to evaporative concentration, and they may, in addition, be fed from above by dilute alpine streams. The alpine waters above 3200 m are only about half as concentrated as those of the montane drainage lakes, perhaps owing mainly to greater precipitation at higher altitudes, as well as to a lesser abundance of the morainic glacial soils commonly found at montane levels.

Troedsson (1955, Chapter 5, p. 99) and Gorham (1958a, p. 176) found that solutions in intimate contact with the mineral soil, as for example those seeping into lysimeters or expressed by pressure, may exhibit ionic concentrations very different from those of normal ground and lake waters, which percolate through the soil mainly along crevices, root channels, worm tracks, stone/soil interfaces, etc. (Tamm and Troedsson, 1957). The expressed or lysimeter waters are much more concentrated, partly because of greater evaporation, but they may also show a very different proportionality among the various ions. Table 14 compares lysimeter drainage and expressed soil solutions with normal ground water and surface waters in middle Sweden (Troedsson, 1955, p. 179), the last two being much alike in this area of shallow and base-deficient soils. The surface waters and ground waters are considerably more dilute than the expressed and lysimeter waters, which are enriched particularly in potassium. Table 14 also compares solutions expressed from the humus layers of woodland soils in the English Lake District and waters of similar chloride content from lakes (Gorham, 1958a, p. 175). The humus layers include both strongly acid and organic *mor* types from heavily leached knolls with semipodzolic soils, and faintly acid *mull* types from flushed brown earths along the edges of streams and lakes (described by Gorham, 1953). The *mor* waters have been more concentrated by evaporation than the *mulls*, as shown by higher chloride concentrations. The ionic compositions of the expressed *mull* and *mor* soil solutions differ not only from one another but also very markedly from those of the lake waters. Potassium is unusually high in both types of soil solution, presumably owing to release from decomposing leaves, which are rich in this element (Ovington, 1956). The *mor* solutions

Table 14. Comparison of Waters Expressed from Soil by Pressure with Ground Waters and Lake Waters from the Same Areas (ppm)

	Ca	Mg	Na	K	HCO_3	SO_4	Cl	NO_3-N
English Lake District								
Expressed soil solution (pH 4.2), from organic *mor* humus layers	1.1	0.9	9.2	5.8	nil	17.5	11.5	0.1
Lake water of similar chloride concentration*	14.5	2.5	5.7	0.9	35.9	14.1	11.3	v. low
Expressed soil solution from circumneutral *mull* humus layers	10.0	1.6	8.7	4.4	8.1	9.6	7.6	7.8
Lake water of similar chloride concentration*	5.3	1.1	4.2	0.4	10.7	7.9	7.4	v. low
Central Sweden (forest land)								
Expressed soil solution	12.8	n.d.	8.7	3.0				
Water seeping into lysimeters	7.0	2.3	5.5	4.7				
Ground waters from wet soils with bog moss	5.6	1.0	3.7	0.7				
Normal ground waters	3.0	1.3	2.8	0.4				
Surface waters	3.2	1.5	5.1	0.8				

* Values calculated from data of Mackereth (1957)

Table 15. Average Composition of Ground Waters and River Waters in Japan (ppm)
(Hanya, 1951)

	Ca	Mg	SO_4	Cl
Ground water	15	5.7	13	15
Smaller river waters	10	3.7	11	10
Larger river waters	11	4.1	13	10

from strongly leached soils are extremely deficient in calcium, as expected. The *mull* solutions are, on the other hand, uncommonly rich in nitrate, because of the activities of bacteria in breaking down organic nitrogen compounds and, in these open and circumneutral soils, oxidizing to nitrate the ammonia that is formed.

Variation in water composition because of different modes of percolation was also observed by Arrhenius (1952) who found that Swedish well waters contain three to four times the amount of soluble salts found in Swedish river waters. Hanya (1951) also found considerably greater amounts of salt in Japanese ground waters than in river waters there (Table 15). Sulfate, however, was not notably higher, perhaps owing to some reduction during percolation through the soil.

ENVIRONMENTAL FACTORS INFLUENCING IONIC COMPOSITION

Climate

The original salt concentration of atmospheric precipitation may depend on the amount of snow or rain falling. Light showery rains are usually much more concentrated than are heavy falls (Sugawara, Oana, and Koyama, 1949; Gorham, 1958a, p. 152), and salinity may sometimes be related to rain intensity (Woodcock and Blanchard, 1955, p. 443). Turner (1955) showed that raindrop salinity may be greatly influenced by conditions below cloud base; he demonstrated a relationship between sodium chloride salinity and raindrop size.

The high contents of calcium, bicarbonate, and especially sulfate in atmospheric precipitation from the arid Ukraine, as compared with precipitation from the moister northwestern parts of the U.S.S.R. (Alekin and Brazhnikova, 1957), are probably due mainly to climatic differences. Were it not for the dry climate of the Ukraine, the soils would not contain such large amounts of easily soluble salts, and dustfall would not be so great either. Similar differences are apparent in Soviet rivers.

Table 16. Comparison of the Observed Concentrations of Chloride in Australian River Waters with Those Calculated from Rainfall Analyses and River Discharges
(V.G. Anderson, 1945)

Catchment area	Mean annual rainfall (ins.) 1897–1914	Discharge / Rainfall	Mean chloride concentration (ppm) In rain 1917–1925	In rivers Calculated	In rivers Observed
Upper Murray	33.19	0.283	1.0	3.5	3.4
O'Shannassy	63.74	0.520	2.7	5.2	4.8
Yarra	44.61	0.311	4.4	14.2	15.3
Barwon	25.33	0.075	8.4	112	134
Helena	28.05	0.034	6.7	197	203

The balance between precipitation and evaporation has a marked effect upon the concentration of surface waters. This is brought out by Table 16, in which the chloride concentration in the rain and the ratio of river discharge to rainfall are seen to act jointly in determining the chloride concentration of Australian river waters (V. G. Anderson, 1945). The good agreement between observed and calculated values suggests little dry fallout or capture of salt particles by vegetation, in marked contrast to results elsewhere. The influence of moisture balance may be seen too in the increased concentrations of chloride (supplied almost entirely by rain) in lowland as compared with upland lakes in the Cairn Gorm mountains of Scotland (Gorham, 1957f), illustrated in Figure 3. The effect of climate is also illustrated by the increase in salt content with decreasing elevation of lakes in central

Sweden. Köhler (1937) records similar results from chloride analyses of dilute Norwegian lakes at very high altitudes.

The influence of moisture balance upon ionic proportions in surface waters was also examined in the discussion of Clarke's views on the ide content of Walker Lake in the Great Basin of the western United States. They suppose that a period of severe aridity dried up the lake, so that its salts were all deposited and covered by sediments when the lake reformed. In other lakes, which apparently did not dry

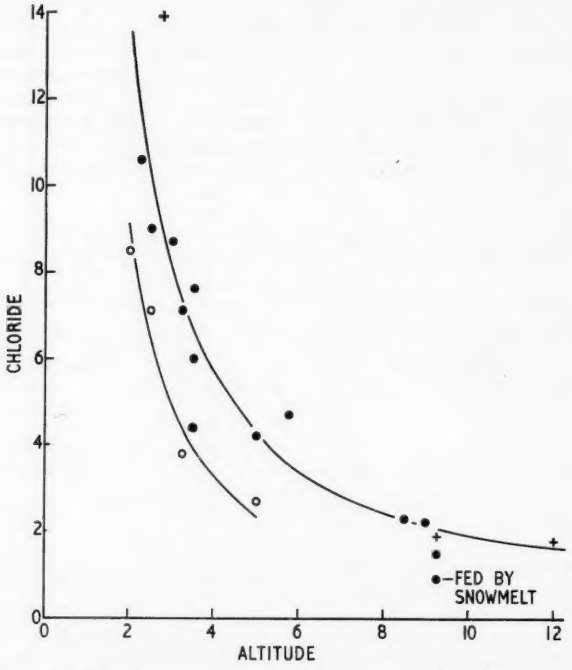

Figure 3. Chloride concentrations of lake waters at varying elevations in the Cairn Gorm region of Scotland. The curve has been fitted free-hand to log/log plots. Solid circles represent small lakes, and open circles large lakes receiving much upland drainage. Crosses represent springs.

importance of climate in determining the chemistry of both river waters and the lake waters of closed drainage basins. Past climates may also affect the composition of lake waters at the present day, insofar as they condition geologic processes. For instance, Gorham (1957c, p. 26) has suggested that the high salt contents of most of the Cheshire meres in England, which are especially rich in chloride and sulfate, reflect the arid conditions in which the underlying Triassic Keuper marls were deposited. An even more extreme situation has been suggested by Broecker and Walton (1959, p. 615) to account for the relatively low chlor- out, much higher chloride concentrations are observed.

Temperature is a climatic factor of the utmost importance, not only in determining moisture balance, but in other ways as well. Snow stored as glacier ice may lose salts through selective washout of intercrystalline brines during the summer melting period (Gorham, 1958h). Strøm (1939) states that lakes fed chiefly by glacier meltwater are notably poor in electrolytes.

According to Cunningham (1941) sea temperature may control the chloride content of sea fog by some sort of evaporation effect; for

air moving over warm water contains many more chloride particles than air moving over cold water. The number of days the air travels over the ocean is important too. Cold weather may also have an indirect effect upon the composition of rain by inducing greater fuel combustion and hence increasing acid air pollution by sulfates and chlorides in winter (*see* Meetham, 1950, p. 362; Dept. of Scientific and Industrial Research, 1955, p. 13; Barrett and Brodin, 1955, Fig. 4, p. 254).

Temperature of course influences the chemistry of surface waters greatly. In cold climates frost splitting of rocks exposes fresh surfaces to weathering. Tropical rivers, which drain strongly weathered soils subjected to a different type of breakdown, are commonly poor in dissolved salts and are peculiar in that a large part of their total salinity is accounted for by silica (Clarke, 1924).

Freezing out of salts may also be of local importance during cold winter weather in high latitudes. Boyd (1959) found that some Alaskan waters become strongly concentrated when most of the water crystallizes as ice. Ion proportions may also be affected, as in some highly concentrated waters from closed basins in Saskatchewan. These waters are normally saturated with sodium sulfate and show freezing out and re-solution of sodium sulfate crystals during cold and warm periods respectively (Rawson and Moore, 1944, p. 149). A similar phenomenon is observable in the Karaboghaz Gulf of the Caspian Sea (Clarke, 1924, p. 168). Thompson and Nelson (1956, p. 237) suggest that some sulfate lakes may have formed through frigid concentration where marine salts were cut off in lagoons or embayments of the sea.

Temperature stratification in lakes can lead to enrichment of bottom waters in salts (Table 13), and sometimes to their depletion in sulfate (through reduction) if the hypolimnion becomes anaerobic (Mortimer, 1941–1942). The types of stratification developed in different parts of the world vary greatly with temperature regime (Hutchinson, 1957, Chapter 7, p. 426).

Another significant climatic factor is wind, whose direction and speed may strongly affect the chemical composition of atmospheric precipitation. Junge (1954; 1956) and Twomey (1953; 1954) reported differences in the chemical constitution of continental and maritime aerosols, and Woodcock (1953, p. 364; *see also* Woodcock and Mordy, 1955, p. 296) showed how sea-salt concentration in the air at cloud level near Hawaii increases with high winds. The range is from about 2 $\mu g/m^3$ with winds of force 1–2 (Beaufort scale) to about 400–500 $\mu g/m^3$ with hurricane winds of force 12. There is a linear relation between the logarithm of sea-salt concentration and wind force on the Beaufort scale; with overland winds such a relationship is not found (Junge, 1954, Fig. 6, p. 329).

Concerning rainfall, Eriksson (1952, p. 289) summarizes the earlier work which describes the influence of wind speed and direction upon chloride concentrations. Barrett and Brodin (1955, p. 257) point to the increased acidity of Swedish precipitation during February-March (mean pH 4.7–4.8), when the predominant air flow during wet spells is from the south and brings air pollution into the country from the great industrial centers of Europe. In May not only does fuel-combustion decline, but also air flow from the north develops in connection with the anticyclone situated over Britain and the North Sea, so that mean pH rises to about 5.7. In summer the blocking anticyclone breaks down, and south or west winds again carry in some acid aerosols, bringing pH values down again to intermediate levels around 5.2. The work of Rossby and Egnér (1955) indicates also that chloride/sodium ratios greater than those of sea water (up to 3.5 as against a sea water ratio of 1.8) characterize precipitation from air streams coming from the south and passing through heavily industrialized regions, while very low ratios (down almost to zero) are typical of precipitation from air masses of northern origin.

The influence of wind direction and wind velocity upon the anionic composition of daily rain samples in the English Lake District has been shown by Gorham (1958a). His Figure 4 (p. 155) illustrates the importance of winds from the southwest in transporting chlorides into the area; the distance between the sea and the sampling point is least in that direction. Southeast winds evidently bring in most of the sulfate, as sulfuric acid, from the great industrial areas lying south and east of the Lake District. The effect of wind speed is much greater upon chlorides than upon sulfates (*see* Gorham, 1958a, Fig. 6, p. 157), presumably because it affects not only transport of sea-salt particles inland but also their entrainment into the air from the sea surface (Woodcock, 1953; p. 364; Woodcock *et al.*, 1953). The magnitude of chloride supply during storms may be

gauged by the fact that a 2-day westerly gale in the English Lake District raised the concentration of sodium in Ennerdale Water by 0.7 ppm, equivalent to a deposit of some 3000 tons of sea water in the drainage basin (Mackereth and Heron, 1954).

The influence of wind direction upon the composition of lysimeter drainage in a New England forest stand has been indicated qualitatively by Joffe (1940, p. 58), who found that a series of peaks and dips in chloride translocation through the A_1 horizon of a gray-brown podzolic soil correlated with rainfall from southwesterly subtropical air masses originating over the sea and from northwesterly continental air masses travelling over the Great Lakes, respectively. In eastern North America, where the prevailing winds are generally seaward, less chloride appears to be carried inland than in western Europe, where the prevailing winds blow toward the land. (*Cf.* Junge and Gustafson, 1957, Fig. 3, p. 166.)

Salt lakes subject to drying may show another effect of wind. Hutton (1958, p. 289) has suggested that certain such lakes in Australia contain far less salt than might be expected from the annual accessions in rain; he postulates considerable losses owing to deflation during periods when the lakes are dry and dusty. Hutchinson (1957, p. 564) notes the possibility of similar losses in the Lahontan basin of western North America.

A different effect of wind upon lakes is its action in breaking down thermal stratification, allowing mixture of enriched bottom water with the more dilute overlying layers. In addition to autumnal cooling, stormy weather with gales and floods frequently assists in bringing an end to stratification of lakes in the English Lake District.

Atmospheric turbulence other than wind may also affect the ionic composition of rain. Junge and Gustafson (1957, p. 167) have invoked large-scale vertical mixing over continents to account for the rapid decrease of chloride in rainfall and in lake and stream waters with increasing distance from the sea coast. This is in contrast to the usual hypothesis that most of the chloride is washed out of the air by rains in the coastal regions, leaving less to be washed out later inland. Over the sea most of the salt is concentrated in the lower part of the troposphere and decreases rapidly above cloud base. Over the land convective mixing is much more active, and the sea-salt particles become dispersed throughout a much taller column of air, so that in the lower layers of the troposphere where most rain develops there are far fewer salt particles available for precipitation. (*Cf. also* Twomey, 1955, p. 84.)

Additional climatic effects are discussed in a following section on the time factor.

Geology

Although considerable attention has been paid to atmospheric accessions of salts, it must be emphasized that soil and rock weathering provides the major supply of ions to natural waters. Durum's (1953) demonstration that a ground-water contribution of only 35 per cent total stream discharge supplied 76 per cent of dissolved solids in the waters of the Saline River in Kansas is a good illustration of this, a situation reasonably typical of sedimentary deposits in the central United States.

Although soils and rocks serve mainly as a source of minerals to ground water, local geology also influences the chemistry of raindrops in the air. The amounts of calcium and potassium contributed to rain by soil dust depend on the nature of the local soils. Rain in the English Lake District, a region of volcanic or slaty rocks, normally contains only about 0.3 ppm of calcium (Gorham, 1955b, p. 236), whereas in a part of western Ireland where limestones are abundant there may be five times this concentration (Gorham, 1957a, p. 2).

The influence of geology upon the composition of lake and river waters has already been dealt with in its general aspects, as outlined by Conway (1942), and numerous local investigations were discussed in detail by Clarke (1924). Clarke also summarized (p. 161) the different geological conditions in which alkaline and saline lakes are formed in closed basins—the former ordinarily characterizing areas of igneous rocks and the latter sedimentary deposits.

Mackereth (1957) and Gorham (1958a, p. 172) showed that on the volcanic rocks making up the central mountains of the English Lake District the waters are very dilute, approximating the composition of rain water plus a little calcium, magnesium, and bicarbonate weathered from the rocks and soils. On the Silurian slates and flags of the southern part of the district soil-derived bicarbonates become much more important, and appreciable amounts of potassium and sulfate are also supplied from the soil, so that the lake waters are considerably more concentrated. On the easily weathered sedimentary rocks of the Triassic and Carboniferous periods, around the periphery of the

Lake District, lake waters are still more concentrated and strongly calcareous. In these waters most of the calcium, about 80 per cent of the magnesium, more than half the potassium, and perhaps one-third the sulfate is supplied by soil weathering. Chloride and sodium in this area seem still to be almost wholly supplied by sea spray carried down in the rain, and their increase in concentration moving outward from the central mountains is probably due to lower rainfall and greater evaporation in the lowlands.

Rock mineralogy may be an important factor in mineral supply. For instance, the volcanic rocks of the central Lake District are not notably poor in lime (Guppy, 1931; 1956), but it is not easily released. Pearsall (1921; 1929), in dividing the English Lakes for ecological purposes into rocky and silted types, recognized this factor.

More frequently of importance is the relative abundance of different elements in various rocks. This effect of geology is clearest in deep-well waters that have had a long time to equilibrate (*see* extensive data summarized by Thresh, Beale, Suckling, and Taylor, 1949; Hem, 1959), but is also evident in surface waters. Limestone influence upon water composition is particularly striking, for where it is present lake and river waters are as a rule greatly enriched in calcium and bicarbonate ions. Figure 1 illustrates this clearly in the case of Swedish lake waters, and Kobayashi (1955) has shown similar enrichment in comparing Japanese rivers draining eruptive volcanic rocks with those draining Paleozoic rocks with limestone. Many alpine lakes in Bavaria and Austria receive calcareous drainage, but owing to the prevalence of dolomite the calcium/magnesium ratio (by weight) of the lake waters averages about 4 (*see* Clarke, 1924, table on p. 100–101) in contrast to an average of about 17 in the Swedish limestone lakes of the Siljan district (Lohammar, 1938; Gorham, 1955a, p. 134). Similarly the calcium/magnesium ratios of Wisconsin lake waters are only about 2; magnesian limestone and dolomite contribute extensively to the soils in their drainage basins (Hutchinson, 1957, p. 555). Triassic substrata in Shropshire also release large quantities of calcium and bicarbonate to the lake waters draining them, which however differ from those of limestone lakes in containing an abundance of other salts, and in their exceptional potassium concentrations (Gorham, 1957d, p. 178).

Table 17. Comparison of Ionic Concentrations in Nova Scotian Lake Waters Draining Granites, Mixed Quartzites and Slates, and Carboniferous Strata

(Gorham, 1957b)

Drainage	Total cations m equiv./l	pH*	Ca	Mg	Na	K	HCO_3	SO_4	Cl
					ppm				
Granite	0.35	4.7	1.0	0.5	5.2	0.4	nil	5.9	7.7
Quartzite and slate	0.29	5.7	2.1	0.4	3.0	0.6	1.8	5.2	4.9
Carboniferous strata	0.37	6.5	3.0	0.6	3.6	0.5	6.1	5.3	5.4

* Arithmetic average of values before aeration

Further examples of the general correlation between geology and water chemistry have been provided by Möller's (1949; 1955) extensive investigations of the dissolved substances in German waters, summarized by concentration maps.

In regard to anions other than bicarbonate, Beauchamp (1953) remarks the importance of sedimentary rocks for sulfate supply and notes that their absence is at least partly responsible for the general dearth of sulfate in East African waters. The influence of sedimentary deposits on the chloride content of inland waters was stressed by Clarke (1924, p. 161), and Eriksson (1955) recorded very high chloride discharges by North American rivers associated with salt domes.

Considerable difference is found in ionic composition of dilute Nova Scotian lake waters on granites, mixed quartzites and slates, or on Carboniferous strata (Table 17). Whereas all the waters are low in salts, those on granite are peculiarly low in lime and are strongly acid. This is counterbalanced by higher sodium levels, since the granite lakes are closest to the sea. The old analyses by Hanamann (Clarke,

1924, table on p. 104), recalculated for Table 18, show that Bohemian stream and river waters draining phyllite, granite, mica schist, basalt, and Cretaceous sedimentary deposits exhibit increasing salt content in that order. Comparable data for some streams in Wyoming and Montana are presented by Swenson (1953). There, waters on granite are dilute (43 ppm total solids) and dominated by calcium and bicarbonate, whereas the waters on limestone are nearly four times richer in solids but are still dominated by the same ions. On gypsum, still more concentrated waters occur (727 ppm total solids), with calcium and sulfate abundant, but the richest waters are those on shale (2020 ppm total solids), in which alkalies and sulfate predominate. Troedsson (1952) has made a somewhat similar comparison of Swedish ground waters, data for which are given in Table 19. These show the importance of Cambro-Silurian limestone in adding calcium and potassium to Swedish waters (*see also* Fig. 1) and the extreme scarcity of these ions in waters draining fluvioglacial sands.

Table 18. Composition of Bohemian Waters Draining Various Geological Substrata (ppm) (Clarke, 1924)

Substratum	Ca	Mg	Na	K	HCO_3	SO_4	Cl
Phyllite	5.7	2.4	5.4	2.1	35.1	3.1	4.9
Granite	7.7	2.3	6.9	3.7	40.3	9.2	4.2
Mica schist	9.3	3.8	8.0	3.1	48.3	9.5	5.4
Basalt	68.8	19.8	21.3	11.0	326.7*	27.2	5.7
Cretaceous rocks	133.4	31.9	20.7	16.4	404.8*	167.0	17.3

* Presumably occurs partly as CO_3

Some influence of soil texture upon water chemistry has been suggested by Gorham (1957d, p. 178) to account for greater salt concentrations in Shropshire lake waters on Triassic sands and gravels than in those on Triassic boulder clay. Presumably the coarse deposits allow deeper percolation and hence more effective leaching of these relatively rich soils, whereas surface runoff is more favored by the boulder clays. Another example is given by Ruttner (1953, p. 57), who remarks that, in the northern approaches to the Alps, springs from lime-poor impermeable soils derived from igneous rocks are richer in calcium than are those from limestone areas in the mountains, where the rain drains rapidly through the permeable rock and has little opportunity to be charged with carbon dioxide in the surface soil.

Subsidiary rocks may have marked local effects upon water composition. In western Ireland the schists of Connemara are occasionally banded with limestone, which may raise the calcium concentration in neighboring waters appreciably (Gorham, 1957e, p. 241). Even more locally, Macan (1950) noted the influence of limestone bands upon water chemistry in different parts of the same small lake, whose inflows ranged in calcium concentration from 3 to 26 ppm. Local mineral springs are often responsible for enrichment of lake waters in sodium chloride (Clarke, 1924, p. 101), while local deposits of gypsum may add calcium sulfate (Clarke 1924, p. 100).

An interesting case of local geological influence has recently been discussed by G. C. Anderson (1958, p. 259), who quotes the views of O. P. Jenkins on the formation of the shallow and strongly saline Hot Lake in north-central Washington, U.S.A., an area of metamorphic rocks, dolomites, and shales. Near the lake are numerous bodies of oxidized pyrite and pyrrhotite, and it has been inferred that during oxidation considerable sulfuric acid was generated and acted upon the dolomitic and other rocks to form soluble magnesium sulfate and the relatively much less soluble calcium sulfate. The magnesium sulfate would be carried into the lake, together with a little calcium sulfate, but the latter would soon precipitate as gypsum and be covered by deposits of epsomite as evaporation led to continual concentration. Mining operations have in fact revealed beneath the lake bottom a 5-meter layer of epsomite, separated from the bedrock by thin layers of gypsum and clay. At present magnesium and sulfate predominate in the strongly saline waters, and calcium is very low proportionately.

Attempts have frequently been made to estimate the relative mobility of elements in the cycle of weathering, by comparing the composition of unweathered rocks with that of

the waters draining away their weathering products

$$\text{(relative mobility} = \frac{\%\text{ of element in water}}{\%\text{ of element in rock}}\text{)}.$$

Smyth (1913) and Polynov (1937, p. 162) were early exponents of the method; later Reiche (1942) and Anderson and Hawkes (1958) applied it to local situations and used the results as evidence that the mobility of different elements varies greatly from one site or rock type to another. Unless a correction for cyclic atmospheric salt is applied, however, erroneous results may be obtained, especially with regard to chlorine and sulfur (Gorham, 1955a, p. 135). In maritime areas sodium and to some extent magnesium mobility may be considerably overestimated. Long-distance aerial transport of dust can also introduce errors, especially in very dilute waters. For instance, the New England precipitation analyses of Junge and Werby (1958) show amounts of sodium, potassium, and calcium that could provide appreciable fractions of the concentrations of these same ions in the New England streams studied by Anderson and Hawkes (1958), especially if the values could be corrected for evaporation. Differential capture of aerial salts by vegetation (Eriksson, 1955; Tamm, 1953, p. 88; Madgwick and Ovington, 1959, p. 17) may also influence such calculations to a degree that is difficult to assess.

Topography

The influence of topography upon water chemistry is often indirect, acting through climatic modification. Effects of altitude upon moisture balance, and hence upon ionic concentration, have been noted for Swedish, Norwegian, and Scottish lake waters. The rise of chloride levels in waters progressively nearer the sea is a more direct effect. Early American work on this subject is cited by Clarke (1924, p. 56), and later data are given for Japanese and English lake waters by Yoshimura (1936) and Gorham (1958a, p. 171). Gorham and Cragg (1960, p. 177) provide some data for bog waters, and Eriksson (1955) shows maps of chloride discharge for Swedish and American rivers. Chloride gradients in rain from various parts of the world have been observed by Leeflang (1938), Hanya (1951, p. 4), Emanuelsson, Eriksson, and Egnér (1954, p. 262), Junge and Gustafson (1957, p. 165), Eriksson (1958, p. 151; 1960, p. 81, 87–89), and Hutton and Leslie (1958, p. 503). Squires and Twomey (1957) give evidence for the effectiveness of orographic rain in washing sea-salt particles from air moving upslope in Hawaii. (*See also* Woodcock, 1953, p. 370.) Eriksson (1958, p. 151) has noted the importance of land form in controlling coastal chloride gradients, which are more pronounced where the land rises steeply from the sea.

Proximity to the sea has also led to temporary marine submergence of many lakes in peculiar topographical situations. Hutchinson (1957, p. 482 ff.) considers several examples where more or less distinct saline water masses persist beneath surface layers of fresh water (ectogenic meromixis). The submergence of much of the Swedish province of Uppland at

TABLE 19. ANALYSES OF SWEDISH GROUND WATERS ON DIFFERENT SUBSTRATA
(Troedsson, 1952)

Substratum	Soil	Specific conductance, micromhos at 20°C	Ca	Mg	Na	K
			ppm			
Sandstone moraine	Iron podzol	16	2.3	0.5	2.6	1.1
Porphyry moraine	Iron podzol	31	2.9	0.3	2.7	1.5
Leptite gneiss moraine	Iron podzol	32	3.0	1.3	2.8	0.4
Fluvioglacial sand	Iron podzol	42	1.7	0.5	5.4	0.4
Dolerite moraine	Iron podzol	46	5.0	0.8	4.2	1.0
Leptite gneiss moraine	Iron or humus podzol	59	5.6	1.0	3.7	0.7
	Wave-washed gravel over glacial clay	115	11.2	3.4	5.3	0.7
Archaean moraine with Cambro-Silurian admixture	Brown earth	420	79.5	4.2	9.3	3.2

the end of the ice age is probably responsible for high chloride levels persisting in the lakes nearest sea level. (*Cf.* Lohammar, 1949.) It is also responsible for the richness of most of the Uppland lakes in calcium and bicarbonate, since layered clays and submarine marls are abundant east of a line from Gävle to the tip of Mälaren (Lunqvist, 1935, p. 293). The lakes around Stockholm, a little to the south, are not so rich (Puke, 1949), presumably because of the lesser deposition of marl there (Gorham, 1955a, p. 142).

Another consequence of the submergence and later uplift of the Fennoscandian land mass was the establishment and subsequent drainage of many aquatic areas, which developed *gyttja* types of sediments rich in various forms of sulfur. This sulfur enrichment has led to extreme acidity in such soils following cultivation, which by improving aeration allows oxidation to build up very high concentrations of sulfuric acid. Wiklander and his associates (1949; 1950a) have studied such *gyttja* soils, and Kivinen (1955) gives several analyses of the water-soluble salts present in them. Somewhat similar types of acid soil are found along the edges of drained swamps in Uganda (Chenery, 1953), where however there is no history of marine submergence. There sulfur reaches the swamps as sodium sulfate, derived from soda-rich minerals in the surrounding hills, which are weathered by sulfuric acid from oxidizing pyrite deposits.

Hutchinson, Wollack, and Setlow (1943) remark that absence of outlet streams may have a marked influence upon the chemistry not only of lake waters but of lake sediments in Indian Tibet. There the closed basins contain waters greatly enriched in magnesium, and in some cases calcium, while their sediments are also high in both these elements. Studies by Böcher (1949) in Greenland and Rawson and Moore (1944, p. 148) in Saskatchewan show a similar contrast between dilute waters in drainage lakes with outlets, and rather saline and alkaline waters in seepage lakes without them. Evaporation is presumably the main concentrating factor in these seepage lakes.

The above examples contrast with the situation in Wisconsin, where seepage lakes are exceptionally dilute, because they collect mainly surface runoff and do not become greatly concentrated by evaporation. A similar phenomenon has been examined by the writer in Epping Forest near London, England. Two small ponds near Baldwin's Hill provide a sharp contrast in location; one (B) is situated on clayey soil in a small valley with a stream running through it, and the other (A) occupies a small hill near by and exhibits no visible inflow or outflow except in time of floods. The seepage pond (A) appears to derive most of its water supply from surface runoff, whereas the drainage pond (B), with a large amount of dissolved silica in its water, undoubtedly receives deep subsoil water via its inflow stream. Analytical data for the two waters are given in Table 20, and it is evident that the deeper soil drainage adds large amounts of salts as well as silica to the water of pond (B); sulfates and alkaline earths are raised most markedly.

TABLE 20. WATER ANALYSES FROM A SEEPAGE AND A DRAINAGE POND IN EPPING FOREST NEAR LONDON, ENGLAND (ppm)

	Ca	Mg	Na	K	HCO_3	SO_4	Cl	SiO_2
(A) Seepage pond	4.0	3.1	4.8	3.5	14.6	8.6	12.1	0.4
(B) Drainage pond	45.4	33.2	24.2	4.7	119.0	144.0	44.4	8.0
B/A	11	11	5.0	1.3	8.2	17	3.7	20

The Cheshire meres in England (Gorham, 1957c) provide another example. Table 21 compares average ionic proportions of seven normal drainage lakes in Cheshire with those of one small and one large seepage lake, and with the ionic proportions of rain from the Lake District to the north, which are probably not greatly different from those in Cheshire rain. It appears that the small lake contains essentially strongly evaporated rain water, whereas in the larger seepage lake a good deal of calcium and magnesium has been added, presumably because of exchange of hydrogen ions in rain for alkaline earth cations adsorbed on soil colloids. Normal Cheshire meres receiving stream drainage from deeper horizons of the soil show this process of calcium addition carried much farther, to the point of producing distinctly alkaline waters. A somewhat similar

picture of progressive calcium addition is shown by the Shropshire meres a little to the south (Gorham, 1957d).

In arid areas the depth to the water table can exert a great influence on water composition. Reservoirs in the plains of Colorado, tapping streams from the mountains, may be divided into alkali and nonalkali waters, the former occupy depressions with high water tables and contain about 2–3 times the calcium and 8–9 times the total mineral material found in the latter (Pennak, 1958). Even in relatively humid regions lesser effects may be discerned, for example Troedsson (1955, p. 179) observed that beneath normal forest in middle Sweden ground water is somewhat poorer in dissolved salts than beneath peaty depressions occupied by *Sphagnum* moss (Table 14).

The nature of the relief frequently determines patterns of sedimentation, which may have marked local effects on water chemistry. In the Swedish province of Dalarna, lakes on Dalälven river sediments are unusually rich in calcium and bicarbonate (Gorham, 1955a, p. 133; *see also* Fig. 1); the source is presumably Siljan limestone drift deposited in late glacial times. The peculiar composition of the waters in Swedish esker lakes, with their uncommonly high proportions of potassium (Gorham, 1955a, p. 134), has also been remarked.

Senescent lakes in Halifax County, Nova Scotia, so situated as to be almost filled by sediments, contain about twice as much calcium in their waters as do normal lakes there (Gorham, 1957b, p. 16). Similarly, from the English Lake District Macan (1950, p. 134) records greater calcium and bicarbonate levels in the waters of tarns in a more advanced stage of colonization by higher plants; the concentrations are also more variable in these later stages. From the Nova Scotian study it also appears that the browner waters in peaty surroundings exhibit low pH values (minimum 4.5), owing to the presence of free sulfuric acid. In this area of base-deficient soils the distribution of peat is controlled largely by topographical factors.

The size and shape of a lake basin may affect the chemistry of its water. From the data on lakes in Scotland (Fig. 3) it is quite apparent that lower chloride concentrations are found in in the larger lakes at a given altitude. This is due not only to the relatively lesser degree of evaporation from these lakes with smaller area/volume ratios, but also to the reception by the larger lakes of a good deal of water from streams draining upland gathering grounds, where evaporation is again less intense. The effect of lake depth upon chloride is illustrated by the data of Yoshimura (1936, p. 348), which show that in Japan, at comparable distances from the sea, the very shallow dammed lakes of Nakakubiki exhibit distinctly higher concentrations than the much deeper crater lakes of Oga. Another example of depth effects is given by Rawson and Moore (1944, p. 149), who find that rates of salinization in some Saskatchewan lakes are related to their mean depths; the rate is less than half as rapid in lakes of 10–15 m mean depth as in those about 1–2 m deep.

The size of bog pools, which determines whether or not they are subject to much drying in summer, affects both concentration and ionic balance in British bog waters (Gorham, 1956b, p. 379). The concentration of sulfuric acid in-

Table 21. Ionic Proportions in Cheshire Meres and Lake District Rain (Gorham, 1957c)

Sample	pH	Total salts m equiv./l	H	Ca	Mg	Na	K	HCO_3	SO_4	Cl
				m equiv.% of total cations or anions						
Rain water	4.5	0.17	26	9	10	52	3.1	nil	42	58
Water from small seepage lake	3.9	0.55	24	16	13	42	5.5	nil	49	51
Water from large seepage lake	4.7	1.05	1	36	20	37	5.7	nil	51	48
Water from seven normal drainage lakes	7.3 to 8.6 (4 samples)	4.53	nil	54	27	17	2.4	48	33	18

creases greatly in the smaller pools during dry periods, owing partly to evaporation but mainly to greater aeration of the peat, and consequent oxidation of sulfur compounds. This increase in acidity leads to exchange of hydrogen ions for metal ions adsorbed on the peat colloids, as mentioned previously, so that sodium/chloride and especially magnesium/chloride ratios increase greatly in the smaller pools. In large permanent bog pools of the Falkland Islands (Gorham and Cragg, 1960, p. 178), some of the sulfate supplied in sea spray may be reduced to sulfide, thus lowering sulfate/chloride ratios below that of sea water; in smaller pools subject to drying, sulfate is considerably enriched, as it is in Britain.

The balance of ions in stratified bottom waters of lakes may be influenced by the shape of the basin, as a comparison of two English lakes shows (Mortimer, 1949). Esthwaite Water, being relatively shallow, becomes anaerobic below the thermocline in summer, while the deep lake Windermere does not. Whereas bottom waters of Windermere exhibit little seasonal change in ionic composition, those of Esthwaite show a marked rise in bicarbonate and fall in sulfate during summer; both effects disappear at the autumnal overturn when isothermal circulation is restored. Hutchinson (1957, p. 675) believes that changes in the slope of the bottom in Linsley Pond, Connecticut, may account for the somewhat irregular vertical distribution of bicarbonate in the lake water during summer stratification.

Some lakes are so large that they contain distinct water masses of different chemical composition, fed by rivers from different parts of the drainage basin, or restricted in circulation and mixing by local topographic barriers or shape. The wide variations in salinity of different parts of the Caspian Sea are classic (*e.g.* Clarke, 1924, p. 168), and Ayers *et al.* (1956) have demonstrated very much smaller but consistent differences in salinity of Lake Huron water masses originating from Lake Superior and Lake Michigan, respectively. Great Slave Lake provides a notable example of isolated and chemically different water masses within one lake (Rawson, 1950). Water from McLeod Bay of Great Slave Lake contains only 23 ppm of total solids and 3 ppm of calcium, although the main lake water contains 160 and 31 ppm respectively. Likewise chloride and sulfate are negligible in McLeod Bay, but the main lake waters contain 12 and 25 ppm respectively. The main lake drains a variety of geological formations, some of them easily weathered and rich in lime and other salts, but McLeod Bay receives only rivers off base-deficient Precambrian rocks, and these develop such a strong flow into the main lake that there is little mixing of the two types of water within McLeod Bay itself. These spatial variations in the chemical composition of water masses within large lakes are analogous to the variations at different points in large rivers owing to chemical differences in successive inflows. Good examples of the latter are given by Clarke (1924, p. 69).

To conclude discussion of the topographical factor, an interesting reverse effect of water chemistry upon lake shape may be mentioned. Hooper (1956, p. 121) noticed that in a series of 27 thermally stratified and calcareous Michigan lakes the alkalinity of the waters is strongly correlated with the mean slope of the basins between the 5- and 25-foot contours ($r = +0.74$). This he interpreted as due to progressively greater calcium carbonate deposition in lakes of higher alkalinity, producing shallow marl benches with a steep marginal drop to deep water.

Biology

Organisms can have a profound influence upon the circulation of elements through natural waters. However, since their distribution is generally conditioned by the environment, it is difficult to find instances where the biotic factor operates independently to control the chemical composition of waters. That is, one can seldom identify primary differences in flora and fauna as responsible for differences in the chemistry of natural waters situated in the same physicochemical environment, except in cases of human interference.

Organisms frequently remove large quantities of the major ionic constituents from waters. Some figures are given in Table 22 for annual mineral uptake (as kilograms per hectare per year) by forests and crops in the temperate zone, and these are compared with data on atmospheric supply and on soil drainage. Such a table does not permit detailed interpretation, but it does show that the quantities of minerals cycled through vegetation are often similar to or greater than the supply by atmospheric precipitation, and may even amount to an appreciable proportion of the quantity carried away by rivers (which of course is removed *after* plants have fulfilled their requirements). Plant demand for calcium and potassium is

very high, but whereas calcium is relatively abundant in most natural waters, potassium is scarce, and a great part of the total available potassium may circulate through vegetation.

Data of Tamm (1953, p. 88; *see also* Gorham, 1955a, Table 2, p. 138) on rain composition before and after percolation through forest canopies indicate an 18-fold enrichment of potassium as compared with 4-fold and 3-fold increases in calcium and sodium respectively. Similar data have been provided by Madgwick and Ovington (1959, p. 17), whose analyses from southeastern England show the following ratios (as kg/ha/yr) of rain-water ions under various forest canopies to rain-water ions in the open: sodium 33/19, potassium 24/3, calcium 24/11, and magnesium 10/<4. The actual potassium circulation through vegetation may therefore be much greater than estimated from Table 22, which takes no account of leaf excretion. The question is further complicated by plant capture of aerosols from dry air, which subsequently release salts when leached by rain (Ingham, 1950; Sugawara, 1951, p. 14; Koyama and Sugawara, 1953; Eriksson, 1955, p. 245). Stenlid (1958) has reviewed the subject of leaf excretion in detail.

Considerable release of mineral elements from forest vegetation also takes place during leaf fall in autumn. Table 22 shows that a large part of the alkalies and alkaline earths taken up by pine or birch forest is returned to the soil in this way (Ovington, 1959; Ovington and Madgwick, 1959); extensive data on nutrient content of forest litter confirm its importance (Stenlid, 1958, p. 625). Curiously, Table 22 indicates that Finnish forest litter is much poorer in potassium than English litter.

Joffe (1933), in an investigation of water percolating through the A_1 horizon of an immature podzol beneath forest, demonstrated a marked rise in pH and specific conductivity during autumn, when leaf fall is greatest. The pH rose from a normal level of 4.8 to as much as 6.4; specific conductivity (a function of total dissolved ions) increased up to four-fold. According to Viro (1955), most of the potassium and roughly half the magnesium in

TABLE 22. CONSUMPTION OF MINERAL ELEMENTS BY VEGETATION, COMPARED WITH AMOUNTS SUPPLIED IN PRECIPITATION AND CARRIED BY RIVERS

(kg/ha/yr)

		Ca	Mg	Na	K	Cl	S
Consumption by pine forest over 55 years, Southeast England[1]		41	7.8	2.4	35		
Litter return		33	5.9	1.6	31		
Net accumulation in trees		8	1.9	0.8	4		
Consumption by birch forest over 55 years, Southeast England[2]		44	5.6	0.6	28		
Litter return		34	4.7	0.5	25		
Net accumulation in trees		10	0.9	0.1	3		
Consumption by 7 English crops[3]	Min.	3	4	2	27	3	3
	Max.	72	29	99	280	93	23
Supply in rainfall, southeast England[4,5]		11	<4	19	3	18	8
River drainage, Europe and North America[3]		82	13	24	9	19	18
Cycled in Finnish forest litter[6]	pine	10	2.0	0.5	2.1		
	spruce	27	4.0	0.3	3.4		
	birch	18	4.1	0.5	3.2		
Supply in precipitation, Kuopio[7]		2.7	0.4	1.0	0.6		
River discharge, Vuoksi system[8]		12.2	3.9	6.0	5.4		

[1] Data of Ovington (1959)
[2] Data of Ovington and Madgwick (1959)
[3] Data from Russell (1950)
[4] Cation data of Madgwick and Ovington (1959)
[5] Anion data from Eriksson (1952), for Rothamsted
[6] Data of Viro (1955)
[7] Data from quarterly appendices to *Tellus*
[8] Data of Viro (1953)

Finnish forest litter is leached out within a year, while only a small part of the calcium is released during this period.

The importance of sulfur release from dead plant material, partly as sulfuric acid, has already been stressed in the discussion of soil weathering. Unfortunately few quantitative data exist for the circulation of sulfur in vegetation. Table 22 gives a range of crop consumption from 3 to 23 kg/ha/yr, and Stenlid (1958, p. 625) quotes a return in forest litter of between 1 and 5 kg/ha annually.

So far only uptake by terrestrial vegetation has been considered. Lund (1957) examined the concentrations of various elements (including calcium, magnesium, potassium, sulfur, and carbon) in the maximum observed phytoplankton crop from a soft-water lake (Windermere) and reported that the amounts of these elements sequestered in phytoplankton amount to no more than a small fraction of the total dissolved in the lake water. However, Beauchamp (1953) believes that in large African lakes the extremely low levels of sulfate in surface waters are due to uptake by plankton, which sediments eventually to the bottom, where its sulfur is trapped in the mud. Such a process may also be important in Swedish seepage lakes (Gorham, 1955a, p. 140). The influence of rooted vegetation in lakes, and especially in small ponds, has not been examined in this way, and complete analyses are scanty. Welch (1935, Table 31, p. 281) quotes some data on the chemical composition of higher plants from Wisconsin lakes.

Animals also withdraw elements from natural waters. Formation of calcareous shells by mollusks may sometimes be of considerable importance, and the lower level of calcium in Lake Tanganyika than in its inflows has been explained by Beauchamp (1939) as at least partly due to molluscan removal.

Return of minerals to lake waters through decomposition of organic materials deposited in the sediments is probably greater in lakes rich in lime and other nutrients. According to Hayes and Anthony (1958, p. 305) the ignition loss of sediments is inversely related to the pH of waters in Nova Scotian and Wisconsin lakes, suggesting greater microbiological breakdown of sedimenting organic detritus in the more fertile lakes.

Apart from direct consumption, plants and animals affect the distribution of major ions in waters indirectly in various ways; perhaps the best known is the precipitation of calcium carbonate marls. Welch (1935, p. 177) gives many examples of marl-forming plants, which by photosynthetic utilization of dissolved carbon dioxide have produced calcium carbonate in excess of the amount soluble. According to Ohle (1934), as quoted by Hutchinson (1957, p. 670), many hard-water lakes may be supersaturated with calcium carbonate, which in surface waters equilibrated with normal atmospheric carbon dioxide pressures would be soluble to the extent of about 65–74 ppm (equivalent to 26–30 ppm of calcium and 79–90 ppm of bicarbonate ion). Much greater concentrations frequently occur in hard surface waters; the excess is presumably in stable colloidal form.

Marl precipitation seems to be favored in waters with low rates of detrital sedimentation. Organic decomposition must be insufficient to generate large quantities of carbon dioxide in the bottom waters and muds; otherwise the marl would redissolve there. In many productive English lake waters very high concentrations of calcium and bicarbonate are present without quantitatively significant precipitation of marl. On the other hand Hooper (1956, p. 122) found that in one group of stratified lakes in Oakland County, Michigan, the formation of marl benches begins to be extensive when bicarbonate alkalinity reaches about 105 ppm, not far above the level reported for calcium carbonate saturation.

In the English Lake District marl lakes are rare, although Sunbiggin Tarn, on the periphery of the district and fed by limestone springs, provides an example. This lake has water with 47 ppm of calcium and 142 ppm of bicarbonate (Holdgate, 1955, p. 403, analyses by Gorham). Dr. J. W. G. Lund (oral communication, 1958) observed incipient precipitation of calcium carbonate in a quarry pool at Brathay, where calcium and bicarbonate concentrations of 37 and 78 ppm respectively were recorded.

Malham Tarn in Yorkshire, south of the Lake District, is another hard-water lake with a complicated history. Table 23 shows analyses (Gorham, 1961) of water from the main inflow and the lake itself, and also the ratio of lake to inflow concentrations. First may be noticed a general reduction of concentration in the lake waters, probably owing to dilution by unevaporated surface runoff; this is illustrated by the chloride ratio of 0.82. Secondly, however, there is a marked reduction of calcium and bicarbonate concentrations; the

ratios are only 0.59 and 0.60 respectively. This may be due partly to accessions of lime-deficient surface runoff, but calcium carbonate precipitation is probably involved as well, since sediment cores reveal marl deposits throughout the greater part of the lake's history. In the recent past, however, much acid peat has been eroding into the lake from an adjacent raised bog; it may react with calcium and bicarbonate or carbonate, adsorb calcium in exchange for replaceable hydrogen ions, and release water and carbon dioxide. (*Cf.* Ohle, 1955b.)

To illustrate the isolating effect of a blanket of peat, unpublished analyses are also included in Table 23 for pools from the surface of the Malham bog. These exhibit the characteristic acidity and calcium deficiency of such waters, completely blocked from drainage percolating mineral soils. In the present instance such soil drainage is alkaline and strongly calcareous.

Table 23. Analyses of Inflow and Surface Waters from Malham Tarn in Yorkshire, and Analyses from Pools on the Adjacent Raised Bog (ppm)

	Ca	Mg	Na	K	HCO_3	SO_4	Cl
Surface water	47.3	2.4	3.4	0.6	143	14.1	4.1
Inflow water	80.0	3.0	3.7	0.7	237	18.3	5.0
Bog pools in wet weather	0.5	0.2	1.6	<0.05	—*	7.8	2.4
Ratio surface/inflow	0.59	0.80	0.92	0.86	0.60	0.77	0.82

* pH 3.9

Another indirect effect of organisms is manifested by the reduction of sulfate to sulfide where biological oxygen consumption has produced anaerobic environments, as for example in the interstitial waters of peats or lake muds, where sulfur may be dissolved as hydrogen sulfide or precipitated as ferrous sulfide. Fixation of sulfur in aquatic sediments is much greater in fertile than in unproductive lakes (Gorham, 1960). Sulfide can also be oxidized again by certain micro-organisms under suitable conditions. Moreover, some sulfate can apparently persist even in completely anaerobic lake muds (Koyama and Sugawara, 1951; Sugawara, Koyama, and Kozawa, 1953, p. 19).

The sulfur cycle has been treated recently by Hutchinson (1957, Chapter 13, p. 753), who gives many examples of the vertical distribution of sulfur compounds in lakes. The work of Mortimer (1941–1942) is particularly interesting in this connection, since in addition to studying lakes he carried out experiments on aerobic and anaerobic mud-water systems in tanks, which provided closed systems and at the same time simulated lake conditions to some degree. Mortimer found that in the aerobic tank sulfate was liberated from the mud into the water, but in the anaerobic tank sulfate in the water was reduced and eventually precipitated as ferrous sulfide on the mud surface. Associated with the final phase of rapid sulfate liberation in the aerobic tank, pH fell, and we may presume (as in the case of the writer's experiments mentioned earlier) that sulfuric acid was being formed by oxidation of sulfides. In analyzing waters from just above the mud in Esthwaite Water during 1940, Mortimer observed a similar situation following the onset of stratification. During the early stages while oxygen remained abundant (although declining rapidly), sulfate increased somewhat in the bottom waters. As anaerobiosis set in, however, sulfate decreased from a maximum of about 9 ppm to 4 ppm in late summer. At the autumnal overturn, when bottom waters were mixed with aerated epilimnion waters much richer in sulfate, sulfate levels just above the mud rose sharply to about 10 ppm.

In shallow bodies of water, reduction to sulfide can probably lower sulfate levels in summer even at the surface (Mann, 1958). Marked decline of sulfate in surface waters of some shallow Uppland lakes during summer (Lohammar, 1938) may be due to such reduction. Possibly the low sulfate levels of some surface lake waters in Wisconsin, Norrland, and Dalarna—much below average levels in rain (*see* Table 5)—may also be caused by reduction to sulfide. In the very saline surface waters of the Sambhar Lake in India (Holland and Christie, 1909), and of Lake Corangamite in Australia (V. G. Anderson, 1945), sulfate levels

are considerably below those expected in waters of oceanic derivation. The authors invoke sulfate reduction, remarking a distinct odor of hydrogen sulfide around the lakes, which at Sambhar becomes "an intolerable stench" when the lake is drying up.

In exceptional circumstances biological reduction may lower sulfate concentrations even in river waters, as shown by analyses of the White Nile above and below the swamps of the "Sudd." In these very extensive swamps, reduction may bring sulfate levels down to 1 ppm or less (Clarke, 1924, p. 109; Talling, 1957, p. 80).

Another indirect effect of organisms upon water chemistry comes from their provision of organic molecules for complexing, and hydrogen ions for weathering and ion exchange with soil colloids. This is important not only in lake muds but also in semiaquatic peats and in normal terrestrial soils.

The effect of a peat blanket upon the chemistry of waters was briefly mentioned in connection with Malham Tarn. In other parts of the British Isles depletion of calcium and bicarbonate, together with increasing acidity, follow upon peat accumulation and isolation of surface waters from the mineral soil (Gorham and Pearsall, 1956; Gorham, 1956a; 1956b, 1957e). Swedish data of Witting (1947; 1948), summarized by Gorham (1955a, Table 3, p. 143), show the same general trends, with pools on deep raised or blanket bog peats apparently deriving their entire mineral supply from atmospheric precipitation.

Human interference is a biotic factor of great importance for water chemistry. Eriksson (1958, p. 156) estimated that human activities might account for 15 per cent of the total chloride and 14 per cent of the total sulfur draining into the ocean, while Hanya (1951, p. 4) reckoned that about 31 per cent of Japanese river chloride is due to human use of salt.

Eriksson (1955, p. 245) believed that potassium fertilizers contribute small but appreciable amounts of chloride to Swedish river waters, and fertilization may be important for sulfate supply to lakes in agricultural districts, since it is commonly an accessory in fertilizers employed for other components (*e.g.*, ammonium sulfate). Such involuntary sulfate fertilization can have profound effects on lake metabolism (Ohle, 1954; 1955a). Bicarbonate status and pH of lake waters can be greatly affected by the use of lime as a soil ameliorant. Addition of NPK fertilizers to lakes, and especially to farm ponds, is frequently carried out to improve fishing, but unfortunately little attention has been paid to its effect upon concentrations of major ions in the water. P. H. Barrett (1957) reports that potassium concentration in surface waters of a small trout lake returned to normal within 1 year of fertilization. Lime and basic slag are now employed in attempts to improve trout fishing in acid bog lakes, but again information upon subsequent long-term changes in pH, alkalinity, and calcium status are generally lacking. When large amounts of basic slag were added to the acid and extremely calcium-deficient waters of two small Nova Scotian lakes, no apparent effect upon levels of calcium or pH could be detected 7 years later (Gorham 1957b, p. 16).

Air pollution is another means by which human beings alter the chemistry of natural waters. Gorham (1955b, 1958a; 1958d) has stressed its significance for the supply of various ions in rain, especially sulfate, and shown that the composition of waters in bog pools and upland tarns on volcanic rocks will thereby be greatly affected. Barrett and Brodin (1955) note the possible influence of dairies, cement works, cellulose industries, and oil-shale plants on the pH of precipitation in Sweden. On a world-wide scale Eriksson (1958, p. 156) estimated that fuel combustion may contribute about 5 per cent of the total sulfur carried by rivers. Junge and Werby (1958, p. 424) calculated that fuel combustion may account for one-third of the non-marine sulfur brought down each year by precipitation.

Irrigation exerts a strong influence on the composition of natural waters in arid districts. On this account Hutchinson (1957, p. 565) found it impossible to use recent analyses in elucidating the chemical evolution of lake waters in the Lahontan Basin of western America. Clarke (1924, p. 69) provided examples of the changes which may occur, citing Headden's data for the Cache de la Poudre River, which originates in the mountains of Colorado and flows out upon the plains. This is initially a dilute and predominantly bicarbonate water with a salinity of 37 ppm. Since it is utilized for irrigation in the arid lowlands, however, it acquires large quantities of easily soluble sulfates, which finally predominate strongly over bicarbonates in the extremely saline waters (1571 ppm) reaching the town of Greeley. Headden's analyses of the Arkansas River in Colorado give a similar picture, but a different one emerges from analyses by Clarke

(1924, p. 158) and others of water from Utah Lake, where irrigation is probably responsible for transforming rather fresh water dominated by sulfate (salinity 306 ppm) into a markedly saline chloride water (salinity about 1300 ppm).

The reverse process of land drainage may also affect water chemistry considerably, resulting chiefly in enrichment of sulfates through oxidation of ferrous sulfide, free sulfur, and organic sulfur in the soil. Ohle (1955a) has noted its significance in fertile North German lakes, where the additional sulfur from this source and from fertilizers has greatly increased the production of hydrogen sulfide in bottom waters during summer stagnation. When sulfuric acid is produced by sulfide oxidation following drainage, the entry of this strong mineral acid into stream and lake waters may neutralize bases and lower bicarbonate concentrations (Ohle, 1953).

In lands that have undergone a period of marine submergence, soil drainage may add chlorides to lake waters through capillary rise of relict marine salts into the upper layers of the soil. Such a situation was revealed through colonization of a Swedish lake shore by plants characteristic of saline soils (Lohammar, 1949). Similar effects may possibly cause relatively high chloride concentrations in the lakes of Uppland nearest sea level (Fig. 1).

Land clearance might also be expected to bring about changes in the chemistry of natural waters. According to Wood (1924) it has increased the rate of salinization of Western Australian reservoirs appreciably; accumulated salts are washed from the soils owing probably to greatly increased drainage through the deeper saline layers of the soil.

Time

The last major factor affecting the chemical composition of natural waters is time. This factor is difficult to examine directly over periods which amount to more than a minute fraction of the age of a lake or river. Most conclusions regarding lake evolution, therefore, rest on the assumption that lakes of similar age but differing characteristics may represent different stages of the same sequence. Such conclusions must be treated with caution.

The likely course of chemical evolution in saline lakes, as construed by Clarke (1924, p. 180) and Hutchinson (1957, p. 566), has already been considered. The general evolutionary tendencies of lakes in less arid regions, however, are considerably more difficult to interpret, since local differences in environment may have much greater influence where the climatic factor is not of such overriding importance. Where leaching of the soil has been intense, for example, and especially where acid peats have accumulated in the drainage basin, lake waters might be expected to become more dilute, acid, and deficient in calcium as the landscape ages; in situations where silting has been pronounced, an enrichment in calcium and other ions might be anticipated.

It seems possible that many lakes were enriched in salts just at the end of the Pleistocene, when pulverized rock deposited by glaciers was releasing large amounts of easily soluble mineral salts. Some of the varves beneath the organic muds of Esthwaite Water in the English Lake District, for example, contain free calcium carbonate, which is absent from the muds themselves (Mackereth, oral communication, 1958). It may therefore be inferred that levels of calcium and bicarbonate in the lake water were higher at the beginning of postglacial time than at any time afterward, even in this silted and rather fertile basin. To investigate release of salts from pulverized rock in the Lake District, Gorham (1958a, p. 165) analyzed a stream water leaching dumps of powdered slate from a quarry and found it contained 2.9 m equiv./l of total salts, and exhibited a pH of 8; most lakes draining similar but weathered substrata had waters with much less than 1 m equiv./l of total salts and circumneutral reaction.

Botanical evidence may sometimes help in reconstructing broad outlines of chemical evolution in lakes. For instance, the composition of the extremely rich diatom floras in the lakes of the English Lake District at or just after the close of the last glaciation reinforces the suggestion that these waters were much richer in calcium than the present lake waters are (Lund, 1959). Indeed Lund suggests that Lake District soils must have been rather calcareous at that time, and probably carried a land flora very different from the present one. (*Cf.* Gorham 1953, p. 148.) Round (1957) found that the diatom stratigraphy of Kentmere in the English Lake District suggests a transition from alkaline to faintly acid waters in the moist early Atlantic period, when peat accumulation in northern England was greatly accelerated. This is in contrast to the hypothesis, also discussed briefly by Round, that evolution of a glacial lake basin commonly involves increasing fertility.

Age, however, was not the only factor involved in the evolution of Kentmere; a climatic shift also strongly influenced lake development. In the history of lake basins it is the rule rather than the exception that changes in other environmental factors complicate the effects of age alone. Thus the time factor, like the biotic factor, can seldom be treated as a wholly independent variable. The effects of Triassic rocks and climate, and of later marine submergence, upon the water chemistry of the Cheshire meres (Gorham, 1957c) provide a good example of the influence of the past upon present conditions, as do also the striking variations in the lake waters of Uppland and Dalarna in Sweden. Effects of air pollution upon natural waters have also increased greatly in the last century or so.

In addition to long-term evolutionary trends, short-term fluctuations are also to be considered, some of a regular recurring nature and others more or less irregular. To various seasonal changes in the chemistry of both rain and lake waters should be added fluctuations over longer periods, as for example those due to less regular climatic cycles of greater duration than 1 year.

The work of Rossby and Egnér (1955) illustrates the effect of such climatic changes on the chemistry of precipitation. These authors observed that between 1949 and 1953 the decreasing ventilation of south Sweden by maritime air masses resulted in lessened deposits of sodium and chloride there, while the increased influence of winds from central Europe led to greater deposition of sulfate.

The effect of time, through climatic change, upon the chemistry of surface waters of some Saskatchewan lakes has been examined by Rawson (1951, p. 671). Here a chain of three lakes in the Prince Albert National Park—Kingsmere, Waskesiu, and Montreal—showed in 1950 a distinct increase of total solids in the downstream direction, with successive values of 169, 188, and 195 ppm in their waters. The lowest of the three, Lake Montreal, drains into Lac la Ronge, which, however, had a lower content of total solids, 179 ppm, owing to reception of considerable drainage from Precambrian rocks. The 1950 data followed several years of normal runoff, but in the drought period of 1930–1936 Lake Waskesiu had little or no outflow, and its total solid concentration rose to 200 ppm. In contrast, Lac la Ronge in 1941 had only 137 ppm, presumably because during the previous 10 years there had been little inflow from the Prince Albert lakes, and most of the drainage had come from Precambrian strata. Thus in the cases of Waskesiu and Lac la Ronge the drought period produced opposite results.

The effect of time, through both seasonal and long-term climatic change, upon the salt content of extremely saline lake waters has also been discussed by Rawson. According to Rawson and Moore (1944) saline lakes in Saskatchewan became much more concentrated during the drought of the 1930's. Data for Great Salt Lake in Utah show two major cycles of increase and decrease over the past century, with the maximum observed salinity about twice the minimum.

Irregular fluctuations in the chemistry of natural waters may also appear within very short periods, owing to the diluting and concentrating effects of wet and dry weather. The chloride concentration of the Saline River, in an arid part of Kansas, varies inversely with stream discharge; the correlation coefficient (r) for a large number of observations is 0.94 (Durum, 1953). The relationship is approximately hyperbolic, with $Cl \times Q = K$, where Cl is the chloride concentration in parts per million, Q is the discharge in cubic feet per second, and K is a constant with a value in this case of 34,500. Chloride discharge is thus about 93 tons per day. Macan (1950) shows the possible extent of such fluctuations in a humid climate, by a series of conductivity and alkalinity analyses from tarns and pools in the English Lake District, while Douglas (1958) provides more extended data for a single small Lake District stream.

Extreme dry- and wet-weather analyses of bog pools and an upland stream given by Gorham (1956b) show striking differences not only in concentration but also in ionic proportions. The size of a body of water is of great importance in this connection, for although large lakes may take years for complete water replacement, the process may take only a few weeks in small lakes (Gorham 1958c, p. 46), and small pools may be flushed out considerably by a single heavy rain. Marked differences in both concentration and ionic proportions observed during repeated sampling of upland tarns in the English Lake District may well be due to such rapid flushing, with rain from different quarters supplying water of differing quality (Mackereth, 1957).

INTERACTIONS AND CONCLUSION

The work of Colby, Hembree, and Rainwater (1956, p. 117) on the streams of the Wind River Basin in Wyoming clearly illustrates the chemical variation of natural waters in response to interactions of partially independent environmental factors. There streams on Precambrian and Paleozoic rocks generally carry dilute waters dominated by calcium and bicarbonate, as do streams on Tertiary volcanic rocks. In contrast, some streams draining Mesozoic strata contain considerable calcium, sulfate, sodium, and bicarbonate. The Bighorn River, which receives all these streams, exhibits marked fluctuations both in ionic concentration and in the composition of its waters, depending on the season and on the amount of precipitation. During minimum flow in winter, concentrations are a good deal higher than during the floods of early summer. However, maximum ion concentrations are observed not during the winter period of minimum flow, when snowmelt provides very dilute runoff, but during precipitation on or ground water discharge from the Mesozoic strata, when the ratios of sodium to calcium and sulfate to bicarbonate increase. The influence of topography is at work here too, insofar as runoff is more rapid in the mountainous Precambrian areas of the basin, and little time is allowed there for solution of salts by rain and snowmelt.

Environmental factors may sometimes exert parallel effects; for instance, low rainfall may act in the same direction as high temperature or easily weathered rocks rich in soluble constituents to produce relatively high ionic concentrations in natural waters. On the other hand, compensating effects may be apparent in certain circumstances. Thus low rainfall, by lessening the degree of leaching, may counterbalance the tendency toward dilute waters where there is a low content of readily soluble ions in the soil. Under a climate humid enough to bring about extensive leaching and peat formation even on lime-rich substrata, acid waters deficient in calcium may be found where under a drier climate distinctly alkaline and calcareous waters might be expected. Such results are exemplified by the acid waters found in parts of the English Pennines, where Carboniferous limestone is overlain in many places by extensive peat deposits (Gorham, 1956b).

The same environmental change may also produce opposite effects in different situations. In the English Pennines wet weather may raise the pH of waters in bog pools, by washing sulfuric acid produced during dry weather out of them and into near-by alkaline streams, whose pH is lowered partly owing to neutralization of their bicarbonate and partly to dilution by rainfall which is also rather acid (Gorham, 1956b, p. 376). Berg (In press) finds that the acidity of streams in the central basin of the Belgian Congo rises during floods for a similar reason.

In any consideration of environmental influence upon water chemistry, it must be remembered that the factors are not wholly independent. For instance, the high salt concentrations of the Cheshire mere waters depend on the richness of the Triassic substrata in salts, determined in turn by Triassic climate and topography, and therefore also by the time factor. Similarly, the increase in chloride concentration of lake waters with decreasing altitude is commonly due to topographic influence upon the climatic factors precipitation and evaporation.

By recognizing that differences in ionic composition so frequently encountered in natural inland waters are commonly due to a complex of interacting environmental variables, we may be impelled to search for situations where differences in water chemistry are clearly due to variation in a single environmental factor. Such a single-factor approach has been most fruitful for soil science (Jenny, 1941) and promises also to be of great value in plant ecology (Major, 1951). Its application to the chemistry of natural waters would seem a logical approach to a subject hitherto investigated not so much for its own sake as for the light it might throw upon related problems.

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Symmetry Concepts in the Structural Analysis of Deformed Rocks

Abstract: Bruno Sander has proposed that the fabric symmetry of a deformed rock reflects the kinematic symmetry of its deformation. In order to place this symmetry principle on a firmer basis, the background of the symmetry theory of fabrics is here reviewed. From very general notions of symmetry as a starting point, fabric symmetry is shown to be a statistical space symmetry consisting of a point group of symmetry operations combined with arbitrary translations in all directions. Where likely restrictions are placed upon the point groups to be expected in homogeneously deformed rocks, the usual types of symmetry observed in fabrics of deformed rocks (namely, spherical, axial, orthorhombic, monoclinic, and triclinic) remain as possible types. The general derivation demonstrates that apart from pseudocrystallographic symmetries defined by some crystallographic fabric elements, no other types of fabric symmetry can be expected in homogeneously deformed rocks.

Attention is drawn to the relevance to the views of Sander of Curie's principles governing the symmetries of "cause" and "effect" in physical phenomena. The features of a deformed rock that define its fabric are found to be a three dimensionally ordered array of discontinuities in structure (lattice planes and lines in crystals, grain boundaries, foliations, lineations, folds, and so on) which may be viewed as the "effect" of deformation. These surfaces and lines of discontinuity in structure are generally sites of surfaces and lines of discontinuity in deformation—implicit in Sander's concept of componental movements—which collectively define the "movement picture" (*Bewegungsbild*) of the deformation. An attempt is made to state more precisely the significance of a movement picture in terms of analysis of local heterogeneities and perturbations in deformation that reflect or are responsible for the development of fabric features in a statistically homogeneously deformed aggregate. The kinds of symmetry observed in tectonite fabrics are found also to be the only kinds possible in movement pictures of homogeneous deformations.

Curie's principles may now be restated in a form directly applicable in the interpretation of tectonite fabrics, thereby amplifying Sander's principle. Without qualification, the most important principle may be restated as follows: whatever the nature of the factors contributing to a deformation may be, the symmetry that is common to them cannot be higher than the symmetry of the deformed fabric, and symmetry elements absent in this fabric must be absent in at least one of the contributing factors.

Resumé: Bruno Sander a émis l'idée que la symétrie de *Gefüge* d'une roche déformée reflète la symétrie cinématique de sa déformation. Afin de placer ce principe de symétrie sur une fondation plus solide, les bases de la théorie de symétrie de *Gefüge* seront passées en revue ici. On partira de notions très générales de symétrie pour montrer ensuite que la symétrie de *Gefüge* est une symétrie spatiale statistique formée d'un groupe d'opérations de symétrie en combinaison avec des translations arbitraries dans toutes les directions. Lorsque des restrictions vraisemblables sont appliquées aux groupes ponctuels l'on peut s'attendre à rencontrer dans les roches déformées de façon homogène, les types habituels de symétrie qui s'observent dans les *Gefüge* de roches déformées (c'est-à-dire sphérique, axiale, orthorhombique, monoclinique et triclinique) demeurent des types possibles. La dérivation générale démontre que, à part les symétries pseudocrystallographiques qui sont définies par certains élements crystallographiques de la *Gefüge*, aucun autre type de symétrie de *Gefüge* ne peut se trouver dans les roches déformees de façon homogène.

Les auteurs attirent l'attention sur le fait que les principes de Curie sur les symétries de "cause" et "d'effet" dans les phénoménes physiques sont appliquables aux idées de Sander. Les caractères d'une roche déformée qui définissent sa *Gefüge* se montrent comme un ensemble, arrangé dans les trois dimensions, de discontinuités de structure (plan et lignes reticulaires dans les cristaux, surface des grains, foliations, linéations, plis, etc.) que l'on peut regarder comme les "effets" de la déformation. Ces surfaces et ces lignes de discontinuités dans la texture se révèlent en général comme des

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lieux de surfaces et de lignes de discontinuité dans la déformation—ce qui est contenu implicitement dans le concept de mouvements composants de Sander—lesquels définissent collectivement "l'image de mouvement" (*Bewegungsbild*) de la déformation. Les auteurs essaient d'exprimer plus précisement le rôle d'une image de mouvement, dans le sens de l'analyse des hétérogénéités et des perturbations locales de la déformation, qui représentent ou qui determinent le dévelopement des caractères de la *Gefüge* dans un aggrégat déformé dé façon statistiquement homogène. Les types de symétrie qui s'observent dans les *Gefüge* des tectonites se montrent aussi comme étant les seules qui soient possibles dans les images de mouvement des déformations homogènes.

Les principes de Curie peuvent maintenant être énoncés de nouveau sous une forme directement applicable à l'interprétation des *Gefüge* des tectonites, et par conséquent, élargissent le principe de Sander. On peut énoncer sans réserve le principe le plus important comme suit: quelque soit la nature des facteurs qui contribuent à une déformation, la symétrie qui leur est commune ne peut être plus élevée que la symétrie de la *Gefüge* déformée; et les éléments de symétrie qui sont absents de cette *Gefüge* doivent être absents d'au moins un de ces facteurs contributifs.

Resumen: Bruno Sander ha propuesto que la simetría *Gefüge* de una roca deformada refleja la simetría cinemática de su deformación. Con el propósito de colocar este principio de simetría en una base más firme, se reseñan aquí los antecedentes de la teoría de la simetría de *Gefüge*. Partiendo de las nociones muy generales de la simetría, se muestra que la simetría *Gefüge* es una simetría espacial estadística que consiste en un grupo de puntos obtenidos de operaciones de simetría, acoplado con traslados arbitrarios en todas las direcciones. En los casos en que se ponen restricciones probables a los grupos de puntos que se esperarían en las rocas homogéneamente deformadas, quedan como tipos posibles de simetría los tipos usuales que se observan en las *Gefüge* de las rocas deformadas (a saber. esférica, axial, ortorrómbica, monoclínica y triclínica). La derivación general demuestra que, aparte de las simetrías pseudocristalográficas definidas por algunos elementos *Gefüge* cristalográficos, no pueden esperarse ningunos otros tipos de simetría *Gefüge* en las rocas homogéneamente deformadas.

Se llama la atención a la aplicabilidad a las nociones de Sander, de los principios de Curie que determinan la "causa" y el "efecto" en los fenómenos físicos. Los rasgos de una roca deformada que definen su *Gefüge*, se encuentra que consisten en un arreglo de discontinuidades en la estructura ordenada en tres dimensiones (planos y líneas del retículo espacial de los cristales, límites de los granos, foliaciones, lineaciones, pliegues, etc.), las cuales pueden considerarse como el "efecto" de la deformación. Estas superficies y líneas de discontinuidad en la estructura generalmente corresponden a los sitios de las superficies y líneas de discontinuidad en la deformación—idea implícita en el concepto de Sander de los movimientos de componentes—las cuales en conjunto definen el "cuadro de movimiento" (*Bewegungsbild*) de la deformación. Se hace un intento de enunciar de manera más precisa el significado de un cuadro de movimiento en los términos del análisis de las heterogeneidades y perturbaciones locales en la deformación que reflejan o son responsables del desarrollo de los ragos de la *Gefüge* en un agregado deformado de manera estadísticamente homogénea. Los tipos de simetría observados en la *Gefüge* de las tectonitas, también se encuentra que son los unicos tipos posibles en los cuadros de movimiento de las deformaciones homogéneas.

Ahora pueden volverse a enunciar los principios de Curie en una forma directamente aplicable a la interpretación de la *Gefüge* de las tectonitas, ampliando así el principio de Sander. Sin restricción, el principio más importante puede volverse a enunciar como sigue: Cualquiera que se la naturaleza de los factores que contribuyen a una deformación, la simetría que es común en ellos (los factores) no puede ser de grado mayor que la de la *Gefüge* deformada; y los elementos de simetría ausentes en esta *Gefüge* deben ser también ausentes en cuando menos uno de los factores contribuyentes.

Zusammenfassung: Bruno Sander hat die Vermutung ausgesprochen, dass die Gefügesymmetrie eines deformierten Gesteins die kinematische Symmetrie seiner Deformation widerspiegelt. Um dieses Symmetrieprinzip auf eine festere Basis zu stellen, wird der Hintergrund der Symmetrietheorie der Gefüge hier besprochen. Die Gefügesymmetrie wird, von ganz allgemeinen Vorstellungen über Symmetrie ausgehend, als statistische Raumsymmetrie dargestellt, die aus einer Punktgruppe von Symmetrieoperationen besteht, die mit willkürlichen, richtungslosen Translationen verbunden sind. Wo im Gefüge deformierter Gesteine den Punktgruppen eventuelle Richtungsbeschränkungen auferlegt sind, was man bei gleichartig deformierten Gesteinen erwarten muss, verbleiben als mögliche Typen die gewöhnlichen Symmetriearten, nämlich die kugelförmige, die achsenförmige, die orthorhombische, die monokline und die trikline Symmetrie. Ganz allgemein geht daraus hervor, dass in gleichmässig deformierten Gesteinen keinerlei andere Typen von Gefügesymmetrie erwartet werden können, abgesehen von pseudokristallographischen Symmetrien, die durch einige kristallographische Gefügeelemente bestimmt sind.

Es wird die Bedeutung von Curies Prinzipien, die das Verhältnis von "Ursache" und "Wirkung" in physikalischen Erscheinungen bestimmen, für die Gesichtspunkte Sanders hervorgehoben. Es hat

sich gezeigt, dass die Merkmale eines deformierten Gesteins, welche sein Gefüge bestimmen, eine dreidimensional ausgerichtete Anordnung von Strukturunterbrechungen sind (Kristallgitterebenen und -linien, Korngrenzen, Lamellierungen, Linear-Strukturen, Falten, u.s.w.) die als die "Folge" von Deformation angesehen werden können. Diese Oberflächen und Unterbrechungslinien in der Struktur fallen im allgemeinen mit den Oberflächen und Unterbrechungslinien in der Deformation zusammen—inbegriffen in Sanders Vorstellung von Komponenten-Bewegung—welche gemeinsam das Bewegungsbild der Deformation bestimmen. Es wird versucht, die Bedeutung eines Bewegungsbild der genauer darzulegen, und zwar in Form einer Analyse von lokalen Ungleichförmigkeiten und Störungen in einer Deformation, die Entwicklung von Gefüge-Eigenschaften in einem statisch gleichartig deformierten Aggregat reflektieren oder dafür verantwortlich sind. Es zeigt sich, dass die Symmetriearten, die in tektonitischen Gefügen beobachtet werden, auch die einzig möglichen Typen in Bewegungsbildern von gleichartigen Deformationen sind.

Curies Prinzipien können nun nochmals in einer Form angeführt werden, die direkt für das Verständnis von tektonitischen Gefügen anwendbar ist; Sanders Prinzip wird dadurch bestärkt. Das wichtigste Prinzip kann ohne Einschränkung nochmals folgendermassen angeführt werden: Was immer die Natur der Faktoren, die zu einer Deformation beitragen, sein mag, die Symmetrie, die bei ihnen gefunden wird, kann nicht höher sein als die Symmetrie des deformierten Gefüges; und Symmetrie-Elemente, die in diesem Gefüge fehlen, müssen in wenigstens einem ser besiteuernden Faktoren fehlen.

М. С. ПЭТЕРСОН И Л. Е. ВЕЙСС

Понятие о Симметрии в Структурном Анализе Деформированных Горных Пород

Резюме: По представлению Бруно Зандера петротектоническая симметрия деформированной горной породы отображает кинематическую симметрию деформации. Для того, чтобы подвести более прочную основу под это принцип симметрии, необходимо ознакомиться с теорией микроструктурного анализа. Начиная с общих понятий о симметрии, петротектоническая симметрия представлена в виде статистической пространственной симметрии, состоящей из системы точек замеров симметрии вместе с произвольной трансляцией во всех направлениях. Там, где возможны ограничения точечных систем в равномерно деформированных горных породах, обычные типы симметрии, которые наблюдаются при микроструктурном анализе деформированных пород (сферическая, осевая, ромбическая, моноклинная, и триклинная), остаются возможными видами симметрии. Общий вывод указывает, что помимо псевдокристаллографических симметрий, определенных какими-либо кристаллографическими микроструктурными элементами, никаких других типов петротектонической симметрии в однородно-деформированных породах ожидать не следует.

Обращается внимание на близость точки зрения Зандера к принципам Кюри в определении характера симметрии «причины» и «следствия» физических явлений. Особенностью деформированной породы, определяющей характер ее микроструктуры, является система нарушений структуры ориентированной в трех измерениях (плоскости и линии кристаллической решетки, границы зерен, сланцеватость, линейная ориентировка, складки и.т.д.), которые можно рассматривать как «следствие» деформации. Эти поверхности и линии нарушения структур обычно являются местом нахождения поверхностей и линий нарушения сплошности во время деформации, подразумеваемой в концепции движения составных частей по Зандеру, сообща определяющих порядок подвижки во время деформации. Делается попытка дать более точные определения значения порядка подвижки путем анализа местных неоднородностей и нарушений во время деформации, отражающих или являющихся причиной появления микроструктурных особенностей в равномерно-деформированном аггрегате. Виды симметрии, наблюдаемые в тектонитах, являются также единственно возможными видами симметрии порядков подвижки однородной деформации.

Принцип Кюри теперь может быть представлен в форме, которая непосредственно применима при интерпретации микроструктур тектонитов и, таким образом,

расширяет принцип Зандера. Без оговорок, главнейший принцип может быть сформулирован следующим образом: какими бы не была сущность факторов, способствующих деформации, симметрия свойственная им не может быть более высшего порядка нежели симметрия деформированной структуры. С другой стороны, элементы симметрии, отсутствующие в этой структуре, должны осуществовать, по меньшей мере, в одном из факторов способствующих образованию этой структуры.

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INTRODUCTION

General

The study of deformed rocks involves the analysis and interpretation of the complex structural features to which the deformation has given rise either directly or by the modification of initial features. Progress in the methods of study of these rocks has been inspired by the pioneering work of Bruno Sander (notably the treatises of 1930; 1948; 1950) and W. Schmidt (1926; 1932). These writings contain, among a wealth of observations on all kinds of geological phenomena, an original approach to problems of strongly deformed rocks based largely upon symmetry arguments.

Thus, many geologists now accept Sander's rule that the symmetry of the structural features of a strongly deformed rock reflects the symmetry of the "movements" involved in the deformation. This symmetry principle has been successfully applied in many studies

of naturally deformed rocks and has been supported by observations on experimentally deformed rocks. An important summary of the experimental evidence and its bearing on symmetry arguments has been published by Turner (1957), and a general discussion of symmetry principles in study of tectonites has been given by Turner and Verhoogen (1960, p. 628–631). However, some geologists, notably E. M. Anderson (1948, p. 121) and Kvale (1945, p. 193; 1946, p. 194; 1953, p. 52, 53), have questioned the validity of symmetry arguments as proposed by Sander.

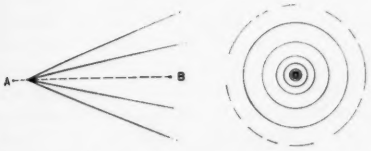

Figure 1. Disturbances with simple symmetry. (a) Symmetry of wake left by ship in motion, (b) Symmetry of ripples on water left by falling stone

The present paper aims to give a general review of the principles of symmetry and their application to the study of deformed rocks in order to clarify the basis of symmetry arguments. Particular attention is given to the following topics:

(1) The nature and description of the symmetry of a rock, the structural features that define this symmetry, and the kinds of symmetry that can occur

(2) The description and symmetry of the mechanical quantities that are relevant to the development of the structural features of deformed rocks, especially the kinematic "movement picture" (*Bewegungsbild*) of Sander

(3) The general rules relating the symmetry of quantities involved in physical phenomena, and their relevance to Sander's symmetry rule.

Role of Symmetry Arguments

By a symmetry argument is meant a deduction concerning the symmetry of an unknown quantity from a knowledge of the symmetry of interrelated quantities. For example, if two quantities can be considered as "cause" and "effect," a knowledge of the symmetry of one enables certain restrictions to be placed upon the possible symmetry of the other.

Such considerations of symmetry permit certain minimum deductions in the study of phenomena for which insufficient information is available for a complete analysis. For this reason, symmetry arguments have been invoked in geology where quantitative information on past physical influences is frequently unavailable and quantitative measurements on the physical properties of the rocks in question have not been made. On the other hand, in physics, where quantitative information on all aspects of a phenomenon can be obtained in the laboratory, symmetry relations are not usually discussed explicitly, although they are implicit in a more complete quantitative description of the phenomenon.

An analogy may be drawn between symmetry arguments and dimensional analysis. Thus, in any equation relating the physical quantities concerned in a given phenomenon, the dimensions must be the same on both sides of the equation, and use of this fact has frequently been made when more knowledge of the quantities is lacking. Similarly, there are general rules governing the symmetry of such quantities. "The connection between the different physical phenomena, as proved by experience, makes it necessary that definite relations must also exist between their special symmetries, in the same way as between their dimensions" (Jaeger, 1920, p. 97). Sander's symmetry rule in structural analysis can therefore be viewed as an application of such symmetry considerations to geological phenomena in order to enable some conclusions to be drawn even though full details are not known.

Simple Examples

The disturbance on the surface of calm water after the passage of a ship in a straight line (Fig. 1a) shows a symmetry from which conclusions can be drawn about the symmetry of the ship's motion. The pattern on the water has a single vertical plane of symmetry *AB*. Therefore, given the premise that the disturbance has been caused by the passage of a ship and that the water was initially still, it may be concluded that the ship's motion had the same vertical plane of symmetry. Also, the polar nature of the symmetry of the disturbance agrees with that of the motion. However, the sense of the motion cannot be deduced on symmetry grounds alone but only from more detailed knowledge of the formation of ship's wakes.

On the other hand, the surface disturbance

resulting from dropping a stone into calm water has axial symmetry about a vertical axis (Fig. 1b). The symmetry of such a pattern has vertical planes of symmetry passing through the axis of symmetry in all azimuths, so there are no polar directions at right angles to the axis of symmetry. This symmetry is therefore incompatible with a cause having polar character in a direction lying in the surface of the water, so that such a disturbance could not have been caused by a body moving in one direction across the surface of the water.

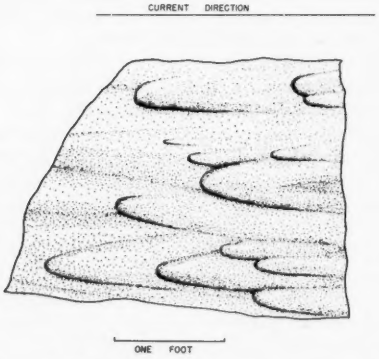

Figure 2. Symmetrical flute casts in graywacke (drawn from photograph by C. M. Gilbert)

These examples show how the study of the symmetry of a disturbance contributes to the knowledge of the symmetry of the motion causing it. In neither case can we distinguish between motions having the same symmetry nor can we determine the velocity without using additional information.

Some Geological Examples

In the interpretation of structures produced in the deposition of sediments by the motion of the depositing medium, geologists frequently make use of symmetry arguments, even if partly subconsciously. For example, from the symmetry of barchans as seen in aerial photographs (*see*, for instance, Finkel, 1959, Pl. 2, p. 615) the direction but not the sense of the wind motion responsible for forming the dunes can be deduced; the wind must be parallel to the vertical plane of symmetry of the dunes, but additional knowledge of barchan formation is needed before its sense can be deduced. The barchans described by Finkel (1959, p. 618) have horns of consistently unequal length. This lack of true bilateral symmetry is attributed by Finkel to the additional influence of the slope of the plane surface across which the barchans are moving. This slope and the accompanying influence of gravity must be included in any symmetry argument concerning the circumstances of the formation of the barchans.

Flute casts on the lower surface of a graywacke bed (Fig. 2) have a symmetry similar to that of barchans, although their origin is different. Again, the direction of current flow, but not its sense, can be deduced from this symmetry; or, if bilateral symmetry is lacking, it may be inferred that influences other than simple current motion are present. The sense of the current can be deduced from other geometrical properties of the structures. Other structures in sedimentary rocks whose symmetry is commonly used in analysis of motion at the depositional interface are ripple marks, cross-bedding, and grain lineations. Lineations defined by aligned elongated grains lacking imbrication, unlike flute casts, are nonpolar structures having two sets of symmetry planes at right angles, so that, on grounds of symmetry, current directions could be either parallel or normal to these structures. Ripple marks produced transverse to currents (Fig. 3a) are asymmetrical in profile, while longitudinal ripple marks (Ten Haaf, 1959, p. 22) are symmetrical in profile (Fig. 3b), consistent with symmetry considerations.

In tectonites, structures at interfaces have similarly been interpreted on symmetry grounds. Thus it may be concluded that the slickensides in a thin layer of mylonite along a fault surface indicate the direction of motion of the fault. However, either sense of sliding is consistent with the symmetry of the slickensides. The procedure, frequently quoted in textbooks, of identifying the sense of the motion with the "smooth" sense for rubbing the hand along the slickensides is therefore subject to error unless justified on other grounds. It has, in fact, been found that the "rough" sense is the sense of motion in experimentally produced slickensides (Paterson, 1958, p. 473–474).

Symmetry Arguments in Physics

The symmetry of the physical properties of crystals and its bearing on their behavior in given situations is of major concern in crystal physics. However, the notions of symmetry can be applied to physical phenomena in

general, and an important paper by Pierre Curie (1894) sets out the principles governing the symmetry arguments that are applicable in physics. These principles have been further expounded by F. M. Jaeger (1920). We shall discuss them in detail later, but the following quotation from Curie serves to illustrate their nature: "When certain causes produce certain effects, the elements of symmetry of the causes must be found again in the effects produced." The similarity to Sander's principle will be noted.

For example, if a simple stress of axial symmetry is applied to a specimen of glass, the glass will acquire the same axial symmetry with respect to its optical refractive index.

We believe that the study of these principles as applied in physics will help to put symmetry arguments in geology on a firmer basis, not only in the analysis of deformed rocks but in the interpretation of many geological phenomena. They may be applied where either the geometrical features of the rocks (as in structural analysis) or the physical properties are being observed. A simple example of the implicit use of a symmetry argument applied to a physical property is found in the inferences about the past magnetic field of the earth from paleomagnetic measurements. Piezoelectric properties have also been observed in some rocks (Parkhomenko, 1958); in consequence, such rocks lack certain symmetry elements (including a center of symmetry), and this lack of symmetry must be sought in the circumstances that gave rise to the piezoelectric properties.

Plan of the Paper

The paper is in two parts. In the first part, a résumé is given of the basic notions of symmetry and the classification of the groups of symmetry elements that can arise. Special attention is thus given to those aspects of symmetry theory that are most relevant to rocks, involving the symmetry of textures (fabrics) and statistical symmetry, as well as to the symmetry of physical quantities. Then follows a review of the general principles relating the symmetries of the quantities involved in physical phenomena. The first part of the paper is therefore self-contained. Its subject matter has application not only in the study of deformed rocks but to other situations in which symmetry arguments may be of value.

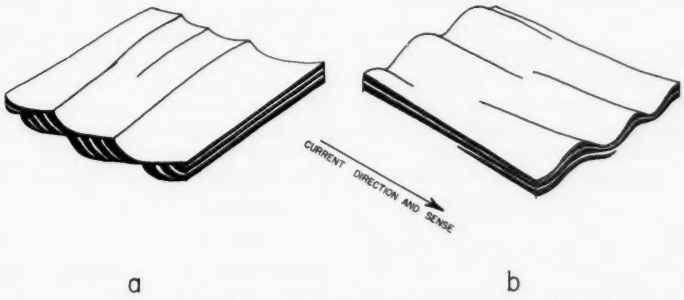

Figure 3. Symmetry of ripple marks. (a) Transverse ripple marks, (b) Longitudinal ripple marks (After ten Haaf, 1959, Dissertation, Univ. Gröningen, p. 22, Fig. 10)

The second part of the paper deals specifically with the application of symmetry arguments to the analysis of deformed rocks. In such an application it is necessary first to have a clear conception of the factors involved. Therefore sections are devoted to the classification of the structural features that are characteristic of deformed rocks and to the quantities involved in the mechanics of the deformation of a body of rock. This is followed by the discussion of the symmetry arguments that can be applied, referring both to Sander's principle and to the implications of the principles of Curie.

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PART I. BASIC SYMMETRY THEORY

General Notions

The symmetry of a body is concerned with the identity of the aspect that it presents when it is observed from different points of view. Thus, depending on the particular aspect concerned, we may speak of the symmetry of a body with respect to its geometrical form, its internal structure, its physical properties, etc.

If the particular features of the body or system under consideration are described with reference to certain co-ordinate axes, the *symmetry* is specified by the assemblage or group[1] of all linear transformations of the co-ordinate axes which leave this description unchanged. Curie (1894) has expressed this as follows: "Let us consider a system defined with the aid of analytical data and of three rectangular coordinate axes, for example. The system will possess a certain symmetry if, in relating it to other rectangular coordinate axes, it is still defined by the same analytical data." Any such transformation of axes, that is, change from one coordinate system to another with respect to which the body appears the same, is called a *symmetry operation*. However, it is usually immaterial whether the symmetry operation is considered as a change of axes relative to the body, which remains fixed, or as an operation on the body relative to fixed axes. The latter point of view is the one usually adopted in crystallography, whereas the former is often more helpful in the study of physical properties.

The study of symmetry itself may thus be regarded both as a branch of co-ordinate geometry and as a branch of group theory. It consists of describing the nature of symmetry operations and of establishing all possible groups of these operations. Particular bodies or systems have symmetry, which is expressed by one or another of these groups of operations. Further, certain classes of bodies or systems may have attributes in common which permit special restrictions to be placed on their possible symmetry groups, as, for example, the restriction to 230 space groups or 32 point groups in crystallography or the more stringent restrictions on the symmetry of optical properties.

The most extensive application of symmetry theory has been in crystallography, and it is therefore with the types of symmetry that exist in crystals that geologists are most familiar. However, the restrictions on the nature of symmetry operations that arise for crystals do not necessarily apply to nonperiodic bodies such as rocks or to physical quantities such as stress. Therefore, we shall first consider symmetry groups from a general point of view and then introduce the restrictions that apply in the study of fabrics and deformation.

Symmetry Groups

General. This section aims to review the nature of symmetry operations and to list the symmetry groups in three dimensions. More detailed discussions may be found in Curie (1884a; 1884b; 1894), Bouasse (1909), Jaeger (1920), Shubnikov (1929; 1955), Hermann (1931), and Nicolle (1950), while a recent treatment of crystallographic symmetry has been given by Buerger (1956).

Symmetry operations. The basic *symmetry operations* considered here are: (1) rotation about an axis by an angle $\frac{2\pi}{n}$ (n = any number ≥ 1); (2) reflection in a plane; and (3) translation parallel to a straight line by a certain distance t. Other symmetry operations arise as combinations of these and include rotation-reflection or mirror-rotation (a combination of rotation and reflection in a plane normal to the rotation axes), inversion about a center (a 2-fold rotation-reflection), rotation-inversion (a combination of rotation and inversion about a point on the rotation axis), screw rotation or helicoidal translation (a combination of translation and rotation about the translation line), and glide reflection (a combination of translation and reflection in a plane containing the translation line). The rotation axes, reflection planes, translation lines and their combinations, which are used in the symmetry operations, are called the *symmetry elements* of the body or system in which they occur.

[1] The term "group" is defined more precisely in the section on symmetry operations.

There are various ways of choosing the minimum set of symmetry operations that is required for a full discussion of symmetry, and different writers have made different choices. In Table 1 are listed the symmetry operations, apart from identity, used in this paper. They are distinguished as symmetry operations of the *second sort* or *first sort*, respectively, according to whether they do or do not change an asymmetric figure into one that can be superposed on its mirror image.

The combination of a pure or "*proper*" *rotation* with a simple operation of the second sort is called an "*improper*" *rotation*. An improper rotation may be regarded as a rotation-reflection or a rotation-inversion according to whether reflection or inversion is chosen as the basic operation of the second sort. These are two different descriptions of the same operation, for which different symbols are used; the difference lies only in the way of resolving the combined operation into component operations. The rotation-inversion description has become conventional in crystallography (International Tables, Henry and Lonsdale, *Editors*, 1952), while the rotation-reflection description has been widely used in other fields. Inversion has been included in Table 1 since it is often useful to distinguish between bodies with a center of symmetry and those without, although the inversion operation is redundant when reflection has been included.

TABLE 1. SYMMETRY OPERATIONS

	First sort	Second sort
Point group operations	1. Proper rotation	1. Reflection 2. Inversion 3. Improper rotation (*either* rotation-reflection *or* rotation-inversion)
Additional operations in space groups	2. Translation 3. Screw rotation	4. Glide reflection

It is convenient at this point to define the term *group* as it is understood in the application of group theory to the study of symmetry. (*See, e.g.*, Hilton, 1903.) An assemblage of operations is said to form a group when it conforms to the following rules:

(1) A combination of any operation with any other operation of the same group, or with itself, is again an operation of the group.

(2) The group contains the operation inverse to every operation of the group and therefore contains also the identity operation.

Particular operations of the group are said to form a *subgroup* when they themselves satisfy the rules defining a group. It follows from the nature of symmetry that all possible symmetry operations for a given body or system form a group, called the symmetry group of the body or system. Among these symmetry groups we distinguish point groups and space groups.

Point groups. If a symmetry group contains no operations possessing a translational component, it is called a *point group*, since its

TABLE 2. POINT GROUPS

10	Schönflies symbol	Nature of symmetry
1	C_n	One n-fold rotation axis
2	D_n	One n-fold axis and n 2-fold axes normal to it ($n \geq 2$)
3	T	Four 3-fold axes and three 2-fold axes, arranged as in a regular tetrahedron
4	O	Four 3-fold axes, three 4-fold axes, and six 2-fold axes, arranged as in a cube
5	P	Six 5-fold axes, ten 3-fold axes, and fifteen 2-fold axes arranged as in a regular icosahedron or pentagonal dodecahedron
6	K_∞	Infinitely many ∞-fold axes, uniformly distributed over all orientations
7	S_{2n}	One $2n$-fold rotation-reflection axis
8	C_{nv}	C_n plus n reflection planes parallel to the n-fold axis ($n \geq 2$)
9	C_{nh}	C_n plus one reflection plane normal to the n-fold axis
10	D_{nd}	D_n plus n reflection planes parallel to the n-fold axis and bisecting the angles between the 2-fold axes ($n=2$ and odd $n \geq 3$)
11	D_{nh}	D_n plus one reflection plane normal to the n-fold axis ($n \geq 2$)
12	T_h	T plus three reflection planes normal to the 2-fold axes
13	T_d	T plus six reflection planes containing the pairs of 3-fold axes
14	O_h	Complete symmetry of the cube
15	P_h	Complete symmetry of the regular pentagonal dodecahedron or icosahedron
16	$K_{\infty h}$	Complete symmetry of the sphere

operations always leave at least one point unmoved. Point groups may therefore include any of the operations listed in the first part of Table 1. Since any finite body cannot possess symmetry elements having translational components but can possess any other symmetry elements, the point groups are sometimes called the *symmetry groups of finite figures*. However, infinitely extended systems may also lack all symmetry elements having translational components and accordingly have symmetry com-

Attention may be drawn to some particular properties of the point groups. When $n=1$, C_n, C_{nh}, and S_{2n} represent, respectively, the simple cases of identity, a single reflection, and the inversion. The first six types of group contain only operations of the first sort and therefore include all the symmetry groups for enantiomorphous bodies. A center of symmetry is present in bodies having symmetry groups C_{nh} and D_{nh} when n is even, D_{nd} and S_{2n} when n is odd, and T_h, O_h, P_h, and $K_{\infty h}$.

Table 3. Crystallographic Point Groups

C_1 1	C_2 2	C_3 3	C_4 4	C_6 6		
	$D_2=V$ 2 2 2	D_3 3 2	D_4 4 2 2	D_6 6 2 2	T 2 3	O 4 3 2
	$S_2=C_i$ $\bar{1}$		S_4 $\bar{4}$	$S_6=C_{3i}$ $\bar{3}$		
	C_{2v} $m\ m\ 2$	C_{3v} $3\ m$	C_{4v} $4\ m\ m$	C_{6v} $6\ m\ m$		
$C_{1h}=C_s$ m	C_{2h} $\frac{2}{m}$	C_{3h} $\bar{6}$	C_{4h} $\frac{4}{m}$	C_{6h} $\frac{6}{m}$		
	$D_{2d}=V_d$ $\bar{4}\ 2\ m$	D_{3d} $\bar{3}\frac{2}{m}$			T_d $\bar{4}\ 3\ m$	
	D_{2h} $\frac{2}{m}\frac{2}{m}\frac{2}{m}$	D_{3h} $\bar{6}\ m\ 2$	D_{4h} $\frac{4}{m}\frac{2}{m}\frac{2}{m}$	D_{6h} $\frac{6}{m}\frac{2}{m}\frac{2}{m}$	T_h $\frac{2}{m}\bar{3}$	O_h $\frac{4}{m}\bar{3}\frac{2}{m}$

pletely described by a point group, as in the example of the electric field around a dipole.

Table 2 lists the possible point groups, with symbols based on those of Schönflies; n is any integer. The point groups may be regarded as comprising all possible combinations of proper and improper rotations, but the operations listed in Table 2 to characterize them are more convenient. The sources quoted above may be consulted for detailed discussion of the point groups and for complete listings of their operations. The symbols C, D, T, O, and P are derived from the names "cyclic," "dihedral," "tetrahedral," "octahedral," and "pentagonal dodecahedral," while the suffixes v, d, and h correspond to "vertical," "diagonal," and "horizontal" (referring to the reflection planes). The symbol S is derived from Spiegel (mirror). The group P is usually called the icosahedral group and is sometimes given the symbol I, but this symbol has also been used for the identity operation.

The group S_{2n} contains neither inversion nor reflection operations when n is even.

It is interesting here to recall the specialization of the point groups of Table 2 which arises in crystallography. It can be shown that Hauy's Law of Rational Indices (a corollary of the periodic nature of crystal structure) implies the restriction to 1-, 2-, 3-, 4-, and 6-fold axes. Thus, by excluding the P, P_h, K_∞, and $K_{\infty h}$ groups and restricting the values of n to 1, 2, 3, 4, and 6, the point groups of Table 2 are reduced to the 32 point groups that characterize the crystal classes. These are listed in Table 3, together with the corresponding Hermann-Mauguin symbols.

Another specialization of the point groups of Table 2 (Curie, 1894), which is especially important in discussing the symmetry of noncrystalline bodies and physical quantities, involves the seven groups that contain an infinitely small rotation (the corresponding symmetry element has been variously called an

infinite-fold rotation axis, an axis of isotropy, or an axis of rotational symmetry). These groups are listed in Table 4. They have been called the *limiting* or *infinite point groups*, there being an infinite number of operations in each group. The first five are the limiting cases ($n \to \infty$) of the point groups of Table 2, having a unique n-fold axis. The two spherical groups K_∞ and $K_{\infty h}$ can be regarded as the limiting cases of those groups which have no unique axes, namely, T, O, P, T_h, T_d, O_h, and P_h, called the endospherical groups by Jaeger (1920, p. 44).

TABLE 4. LIMITING OR INFINITE POINT GROUPS

No.	Schönflies symbol	Nature of symmetry	Characteristic example
1	C_∞	One ∞-fold axis	Rotating cone
2	D_∞	One ∞-fold axis with 2-fold axes normal to it	Twisted cylinder
3	$C_{\infty v}$	One ∞-fold axis with reflection planes parallel to it	Cone
4	$C_{\infty h}$	One ∞-fold axis with a reflection plane normal to it	Rotating cylinder
5	$D_{\infty h}$	D_∞ plus reflection planes normal and parallel to the ∞-fold axis	Cylinder
6	K_∞	Infinitely many ∞-fold axes, uniformly distributed over all orientations	Sphere filled with optically active liquid
7	$K_{\infty h}$	K_∞ plus reflection planes of all orientations	Sphere

The examples given in the last column of Table 4 and depicted in Figure 4 illustrate the symmetry of the limiting point groups. The arrangement in Figure 4 shows the subgrouping; each point group is a subgroup of those above it. A center of symmetry is possessed only by systems having the symmetry $C_{\infty h}$, $D_{\infty h}$, and $K_{\infty h}$. Groups C_∞, D_∞, and K_∞ contain no operations of the second sort and can represent the symmetry of enantiomorphous bodies. Bodies with physical properties of spherical symmetry K_∞ are sometimes said to be "pseudoisotropic," since, although their properties are the same when measured in any direction (conforming to the usual definition of isotropy), they lack the reflection planes and center of symmetry that are often taken to be implicit in the definition of "complete" isotropy. There is therefore the possibility of confusion in the use of the term "isotropy" as applied to enantiomorphous bodies.

The following are physical examples having limiting point symmetry:

C_∞ : a fluid in helical motion (as in a cyclone)
D_∞ : the optical activity of a quartz crystal
$C_{\infty v}$: a homogeneous electric field
$C_{\infty h}$: a homogeneous magnetic field
$D_{\infty h}$: the optical indicatrix of a uniaxial crystal
K_∞ : the optical activity of a sphere of optically active liquid
$K_{\infty h}$: the optical indicatrix of a cubic crystal.

Space groups. In general, the symmetry of an infinitely extended body or system may include any of the symmetry operations listed in Table 1. When the symmetry includes operations with translational components, it is said to comprise a *space group*. Since a translation can be repeated indefinitely without bringing all points of a body back to their original positions, a space group necessarily characterizes an infinitely extended body or system. For example, in speaking of the space group of a crystal structure, reference is being made to the symmetry of a configuration which is envisaged as extending indefinitely in all directions; no account is being taken of the boundaries (regular or otherwise) that a real crystal may possess.

The translations or translational components of the operations of a space group may or may not include infinitesimal displacements. If the least translational component in a given direction is of finite magnitude, the space group can contain only symmetry operations the magnitude of whose translational components in this direction are integral multiples of the minimum displacement. This follows from the properties of a group given above. Thus, if the minimum translational component that any operation can have is of magnitude t and we suppose that another operation of the group has a translational component, in the same direction, of magnitude $nt+\Delta t$ (where n is any integer and Δt is smaller than t), then the combination of the latter operation with the inverse of an operation having the translational component nt will give an operation having the translational component Δt, which must also be an operation of the group; but this is forbidden. Therefore, the magnitudes of the translational

components, in the given direction, of any operations of the group are limited to the values *nt*. On the other hand, it follows by a similar argument that if any operation can have an infinitesimal translational component in a given direction then the translational components of other operations in the groups can have any value in this direction. In the former situation, the equivalent points[2] in a system having the symmetry of such a space group are discretely spaced in the given direction; in the latter situation, the equivalent points form a continuum in the given direction.

With the aid of these notions, we shall now consider special types of space groups for three-dimensional systems:

(1) *The 230 crystallographic space groups* (the Schönflies or Federov groups) apply to the three-dimensional systems (the *Diskontinuum* of Niggli, 1919) in which the equivalent points are discretely spaced in all three dimensions, corresponding to the requirement that the least translational components of the symmetry operations be finite in all directions. The only rotations or rotational components compatible with this requirement are the 1-, 2-, 3-, 4-, and 6-fold ones. (*See*, for example, Buerger, 1956, p. 33.) Since operation with a 1-, 2-, 3-, 4-, or 6-fold screw axis or a glide plane leads to equivalent points among which a certain fraction are translationally equivalent,[3] it follows that the crystallographic space groups always contain a subgroup of pure translations—that is, one of the translation groups of the 14 Bravais lattices. The magnitude of the smallest pure translation in a given direction is often called the "identity distance" in that direction; it is always finite in systems whose symmetry conforms to one of the crystallographic space groups, and these systems are therefore periodic in three dimensions.

(2) *The space groups of a three-dimensional continuum* are involved when the symmetry operations can have infinitesimal translational

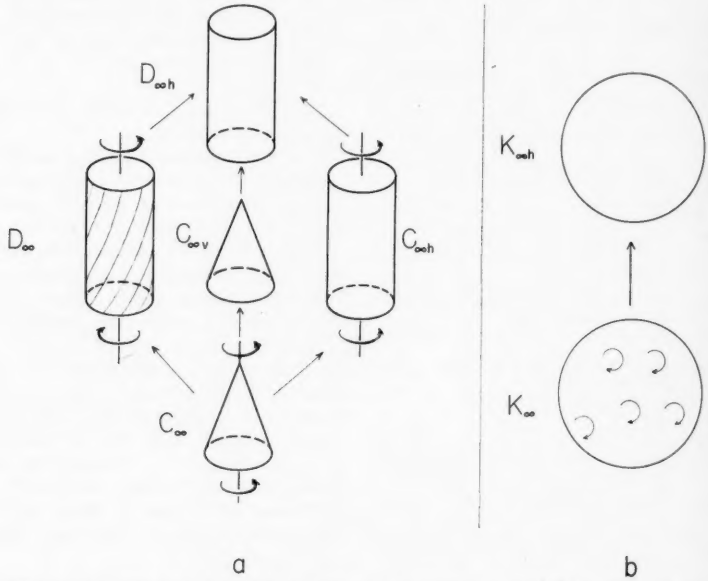

Figure 4. Examples of systems having limiting groups of symmetry, with (a) a unique axis; (b) spherical symmetry

[2] The equivalent points are those that are related to a single point by all the operations of the group.

[3] Translationally equivalent points are those that can be superposed by pure translations. The operation of an *n*-fold screw axis gives no translationally equivalent points when *n* is irrational (Bernal, 1959).

components in any direction. These are the groups "whose symmetry elements generate from a point a system of equivalent points which continuously fill space" (Shubnikov, 1929).[4] As a consequence of permitting infinitesimal translational components, screw axes become continuous screws, and glide planes become identical with pure reflection planes. Therefore, the only symmetry operations with translational components to be considered are the pure, arbitrary translation and the continuous screw. The space groups of a three-dimensional continuum have been derived by Shubnikov (1929; see also the comments by Heesch, 1930, on the general applicability of this derivation); he lists them in three categories:

(a) The combination of any point group of Table 2 with arbitrary translations in all directions; for example, a homogeneous electric field has the symmetry of the combination of point group $C_{\infty v}$ with arbitrary translations in all directions. The space groups in this category are of special importance for fabrics (textures).

(b) The combination of the C_n and D_n point groups (including D_1 for convenience of description) with continuous screw axes parallel to the n-fold axis and arbitrary translations in all directions normal to it; for example, a uniaxial, optically active crystalline medium has such a symmetry (with the D_1 point group) with respect to its optical activity.

(c) The group containing, apart from identity, only infinitely many equivalent continuous screw axes passing through every point in space in all directions; for example, a pseudoisotropic, optically active medium has this symmetry with respect to its optical activity.

3. Various other types of space groups are concerned when the minimum translational component of any symmetry operation is specified to be finite in only one or two directions (Heesch, 1930; Shubnikov, 1930).

Every space group is said to be *isomorphous* (or isogonal) with the point group that is formed when all translational components of its symmetry operations are neglected. If the only operations of the space group having translational components are pure translations, then the isomorphous point group is a subgroup of the space group. Under other conditions, the isomorphous point group is not a subgroup of the space group; for example, the point group of quartz is not a subgroup of its space group, since the point group contains a pure 3-fold rotation whereas the space group contains only 3-fold screw rotations.

When observations are made on a scale that is large compared with the minimum pure translation (identity distance), the body or system will appear to have the spatial symmetry of a continuum, specified by the combination of arbitrary translations in all directions with the point group isomorphous to the space group. This is the basis for treating a crystalline medium as a homogeneous continuum with the point symmetry of one of the crystal classes, as is usual in crystal physics.

Statistical symmetry. So far we have considered only systems that are perfect or ideal in their symmetry; that is, it has been assumed that a symmetry operation brings the whole system into an aspect that is identical to its original aspect in all details. Real bodies or systems may approximate this ideal in various degrees. For example, crystals may approximate ideally symmetrical bodies very closely. On the other hand, in systems such as aggregates of crystals, noncrystalline arrangements of particles, and fields of turbulent motion, whereas there can be operations that bring the configuration or properties of the system into an aspect that is identical with respect to the average of certain features, there may be no operations (except identity) that lead to an aspect that is indistinguishable in all details from the original. In this case, such systems are said to possess statistical symmetry.

A *statistical symmetry operation* may be defined as follows: If we associate with each point in the system or body the average of the relevant features within a certain finite volume surrounding the point and these averages are described with reference to certain co-ordinate axes, a statistical symmetry operation is any linear transformation of the co-ordinate system that leaves the description unchanged within a certain error. In general, the error will be smaller when the averages are taken over larger-volume elements. The types of symmetry operations that may arise are the same as listed in Table 1, and they form the same groups. The notion of statistical symmetry may also encompass time averages. Thus, strictly speaking, the symmetry of crystal structure that is usually discussed is that of the time average of the instantaneous configurations

[4] This statement may be taken as defining a continuum, at least insofar as its symmetry is concerned. Alternatively, we may define a continuum as a medium or region in which every volume element, no matter how small, contains material characteristic of the medium or points of the region. (*Cf.* Lindsay, 1951, p. 279.)

that occur during the thermal vibration of the atoms.

There is a distinction between statistical symmetry, as just defined, and the pseudo-symmetry that is often mentioned in crystallography. The latter refers to operations that bring a perfect system into a *nearly* identical aspect in *all* details; for example, a monoclinic crystal in which the crystal axes are nearly mutually perpendicular is said to be pseudo-orthorhombic. It follows that statistical pseudo-symmetry may also arise.

There has been comparatively little discussion of statistical symmetry in the literature of symmetry, although it is always implied that some imperfection can be allowed for in discussing the symmetry of real bodies, for example, the form of flowers or animals. However, Weissenberg and Hermann have written papers that are especially relevant to the statistical symmetry of textures or fabrics.

Weissenberg's papers (1922; 1924) are concerned with the "statistical anisotropy" of bodies and its X-ray determination, the emphasis being on the fabrics of natural products such as fibers, muscle, and hair, and of polycrystalline aggregates (especially metals) that have a preferred orientation of crystallites as a result of plastic deformation, casting, electrodeposition, etc. The symmetry of the "anisotropy" (fabric) of such bodies is viewed as a statistical property, related to the statistical order in the arrangement of their constituent particles or crystallites. The arrangement of crystallites is represented by the projection of their crystallographic directions on a sphere, giving a density distribution on the sphere from which the statistical symmetry is evident.

Hermann's paper (1931) is concerned mainly with the symmetry of mesomorphic phases or "liquid crystals," but the theory is of general interest. In particular, the discussion of the so-called "translation types for amorphous bodies" is relevant to the fabric of rocks. Hermann expounds some rather complicated notions about statistical symmetry under two main headings, "Statistical Translations" and "Statistical Symmetry Operations," but these will not be discussed here.

Symmetry of Fabrics (Textures)

The applications of symmetry theory in structural crystallography are well known. Here we shall discuss the symmetry of the "structure" of noncrystals. By analogy with crystal structure, the "structure" of a noncrystal is the internal geometrical configuration of its elementary parts and of any characteristic features to which the arrangement of these parts gives rise. Examples of such internal geometrical configurations are seen in the preferred orientation of individual crystals in deformed metals or rocks, the slag stringers in wrought iron, the foliations and lineations in deformed rocks, and the fibrous character of wood and similar materials. This "structure" or configuration of particles (crystalline or otherwise) in an aggregate has been widely termed *texture* in physical and metallurgical literature. The geological term *fabric* (*Gefüge*) is synonymous with "texture" in this usage but is preferred in geology because "texture" has other meanings. Therefore, the term "fabric" will be substituted for "texture" when referring to non-geological literature. Sometimes the term "texture" or "fabric" is used for the body itself, but it should be borne in mind that, strictly speaking, the fabric refers to the internal configuration of the body. Moreover, since no account is to be taken of any boundaries, a fabric can be considered to be of infinite extent.

In the ultimate analysis, all bodies are heterogeneous, and it is just the configuration of this heterogeneity that constitutes the fabric. (*Cf.* the configuration of atoms or electron density in the case of crystals.) On a coarser scale, however, such bodies may appear to be homogeneous. It is therefore of primary importance to state clearly the notions of homogeneity of fabric in relation to scale, since any discussion of symmetry of fabric presupposes some degree of homogeneity or regularity.

"A body is called homogeneous when any two equal, similar parts of it, with corresponding lines parallel and turned towards the same parts, are indistinguishable from one another by any difference in quality" (Kelvin and Tait, 1883, § 675). This definition is to be applied in practice with reference to a certain scale of observation, which in the case of fabric means choosing the "parts" large enough to give a representative sample of the heterogeneities that make up the fabric. Thus we may define a *statistical homogeneity* as follows: the fabric of a body is statistically homogeneous on a certain scale when the average of the internal configuration in any volume element is the same for all volume elements with dimensions not smaller than the scale of consideration.

Homogeneity can also be expressed in terms of translational symmetry, as Weissenberg

(1931) has pointed out: "homogeneity can be defined as a symmetry property of the object, expressing that, for each coordinate system with a given origin, one can derive identical descriptions by translations in all three space dimensions." In the case of crystals, there is a minimum value for such translations in any given direction, which is set by the size of the region of nonidentity that is being repeated periodically and which is an expression of the discrete nature of matter. The homogeneity of a crystal therefore refers to the periodic repetition of a certain finite, inhomogeneous region, the region of nonidentity. On the other hand, a fabric that is statistically homogeneous on a given scale has the same statistical description for all translations (including infinitesimal translation) of a reference co-ordinate system. There is no lower limit to these translations because the volume elements over which the average is to be taken may overlap to any extent. Therefore, *any fabric that is statistically homogeneous on a certain scale can be regarded as a homogeneous continuum with respect to its statistical symmetry on this scale.*

It follows that the possible symmetry groups of a statistically homogeneous fabric are those of the first category of Shubnikov's space groups of a continuum, mentioned above. The restriction to the first category is implied in the definition of homogeneity, which requires the presence of the group of all pure translations. With every point in the continuum is to be associated the average configuration of the fabric in the finite volume surrounding it; this will have a symmetry which will be the point symmetry isomorphous with the space group of the continuum. The consequences of any "screw character," such as might be implied by the presence of optical activity in the body, can be included in the specification of this point symmetry, provided the identity period of the screw is smaller than the scale of consideration; if not, the fabric cannot be regarded as homogeneous. Therefore, in summary, the possible *space groups of a statistically homogeneous fabric* comprise the combinations of the point groups of Table 2 with all possible translations.

Such a rationale of the symmetry of fabrics corresponds to that given by Shubnikov (1955) in his discussion of "textures," and foreshadowed at the end of his 1929 paper: "under the assumptions that space contains infinitely many very small material particles which are nearly the same, which have the symmetry of a point group G, and which are in parallel arrangement with approximately uniform density in space, such a statistically homogeneous space has symmetry $\{G\ T'''\}$." Here T''' is a symbol for infinitely many axes of arbitrary translation in all directions. Shubnikov (1955) defines a "texture" (fabric) as follows: "by the term *texture* we mean all homogeneous bodies without a lattice and formed of a multitude of elementary bodies of nondefined physical nature, oriented in space in a way determined according to the laws of symmetry." Regarding the symmetry of such a texture, "if the initial particle possesses a symmetry corresponding to a point group G, one obtains by translation T a texture of space group $(G)\ T$, where T comprises "all the translations possible in magnitude and direction, including the infinitely small." Shubnikov then discusses the special importance of fabrics based on the limiting point groups (Table 4), including the fabrics of aggregates of needle-shaped and lamellar particles. Shubnikov's concepts of fabrics and their symmetry have been extensively applied by himself and other Russian writers in the discussion of piezoelectric polycrystalline materials (for example, Zheludev, 1957a; Bazhenov, 1957), while Shafranovskii (1956) has also applied them in structural petrology.[5]

In the earlier work of Weissenberg (1922; 1924) on the symmetry of fabrics, only the symmetry groups that include the inversion operation are considered, for two reasons: (1) if the individual crystallites in an aggregate have a center of symmetry, the aggregate must have a center of symmetry; (2) in any case, most methods of investigation (especially X-ray methods, with which Weissenberg was particularly concerned) depend on centrosymmetric properties and are not capable of revealing the absence of a center of symmetry. Therefore, Weissenberg's list of possible point groups (his Table 3) contains only the groups with an inversion from Table 2 of the present

[5] The notation for space groups of fabrics used by Shubnikov and the Russian writers who follow him is based on symbols for the point groups which differ from those of Schönflies and from the international crystallographic symbols of Hermann and Mauguin. It is suggested here that the space group of a fabric can be more conveniently represented by the Schönflies point group symbol to which is added the superscript T; for example, the symmetry of the "fiber texture" of a drawn metal wire would be represented by $D_{\infty h}^{T}$. Hermann-Mauguin symbols are unsuitable since they are based on crystallographic sets of axes.

paper and comprises eight so-called "anisotropy classes" which may be represented in Schönflies symbols as follows (in the order given in Weissenberg's table):

C_{nh} (n even); D_{nh} (n odd); T_h; D_{nh} (n even)
O_h , P_h ; $D_{\infty h}$; $K_{\infty h}$.

Weissenberg omits the class S_{2n} (n odd), which includes the triclinic group with a center of symmetry ($S_2 = \bar{1}$); this should therefore be added to his list.

Weissenberg distinguishes two cases of spatial symmetry. In the simpler case of a homogeneous phase (corresponding to our statistically homogeneous fabric), the symmetry (or "anisotropy") at all points in the aggregate is statistically the same, so that the point group of the "anisotropy class" is a sufficient description of the symmetry (the arbitrary translations in all directions are implied). In the case of heterogeneous phases, the "anisotropy" must, in general, be determined at all points, and the specification of the symmetry includes "in addition to the anisotropy at a point, the nature of its spatial variation." Weissenberg also discusses the particular crystallite arrangements for a given "anisotropy class."

Symmetry of Physical Quantities

Notions of symmetry may be applied not only to geometrical configurations but also to physical quantities. In order to discuss the symmetry of a physical quantity, it is necessary to have an adequate description or representation of it. (*Cf.* "the 'image' of a physical state" of Jaeger, 1920, p. 93.) This may be a geometrical representation, for example the optical indicatrix representing the refractive index of a medium, or the strain ellipsoid representing a homogeneous strain; or it may be a purely mathematical description of the quantity, for example the law of variation with direction of the refractive index and its principal values, or the strain tensor. If the description of the quantity is referred to certain co-ordinate axes, its symmetry is given by the group of all linear transformations of the co-ordinate systems that leave this description unchanged.

"The symmetry of a quantity is defined by its invariance with respect to a group of transformations of co-ordinates. We say, in particular, that a quantity is isotropic or anisotropic according as it is invariant or not with respect to a group which contains all rotations" (Weissenberg, 1935, p. 103). Further, a quantity is homogeneous if it is invariant under all translations in all directions. In other words, a quantity is isotropic if any measure of it is the same in all directions, and it is homogeneous if its specification with respect to given co-ordinate axes is the same at all points (except for change of origin). As noted above, isotropic quantities fall into two classes, depending on whether their point symmetry is K_∞ or $K_{\infty h}$.

Physical quantities are of two kinds, which may be distinguished as *physical properties* and *physical fields*[6] (Nye, 1957, p. 91). Examples of physical properties are the optical properties, elasticity, and strength of crystalline and other bodies; physical fields include stress, strain, and electric and magnetic fields. The former can be specified only with reference to particular bodies, while the characteristics of the latter are independent of particular bodies or substances. In particular, the orientation of the symmetry elements of a physical property is related to that of the symmetry elements of the "structure" or internal configuration of the body, but there is no such limitation on the symmetry of a physical field.

A body may be said to possess a certain symmetry with respect to a given physical property, this being the symmetry of the physical property. It is a general rule that the physical property may possess more symmetry elements than the internal configuration or "structure" of the body, but not less. A given physical property may possess certain symmetry elements in all cases; for example, all bodies have a center of symmetry with respect to optical refractive properties. Thus, there is a minimum symmetry for a given physical property, to which can be added other symmetry elements in particular circumstances.

A physical field can, in general, be regarded as a continuum and can therefore have the symmetry of any of the groups for a continuum. As in the case of physical properties, however, there may be a minimum symmetry that is characteristic of a given type of physical field under all circumstances, to which additional symmetry elements can be added in particular cases.

Since both types of physical quantities frequently have rotational symmetry about one or more axes (infinite-fold axes), the limiting point groups are especially important. Examples of these have already been given above. However, if the quantity is homogeneous, the complete specification of its symmetry is a

[6] Sander (1948, p. 7, 31; 1950, p. 5) uses this term. The concept is apparently included also in his definition of "functional" fabric (1948, p. 6, 7).

space group—namely, the combination of a point group with all arbitrary translations.

There are two important cases of inhomogeneous physical quantities. The first consists of a property or field that varies in a simple regular manner in space (*cf.* Love's "curvilinear aeleotropy," 1927, p. 161), for example, the magnetic field around a bar magnet. The symmetry of the whole field is completely specified by a point group. The second case is that of statistically homogeneous quantities, which have statistical symmetry analogous to that of statistically homogeneous fabrics. Examples of these are found in turbulent motion in fluids, the strain field in a deformed polycrystalline aggregate, or the physical properties of an aggregate.

Many physical quantities can be represented by tensors of appropriate rank, including scalars and vectors (Nye, 1957). The symmetry of the quantity can then be specified by the group of all linear co-ordinate transformations under which the components of the tensor remain invariant. This has led Shubnikov (1949; 1951) to determine and tabulate the point groups of scalars, vectors, and second-rank tensors. Zheludev (1957b) has slightly extended this list and has shown that there are altogether 17 possible groups of symmetry for these quantities and their combinations. Zheludev (1957c) has further pointed out that these groups comprise all possible groups of symmetry that a homogeneous, continuous isotropic medium can take on when under the influence of fields described by such quantities.

Symmetry Relationships in Physical Phenomena

The previous sections have outlined the ways of describing the symmetry of the internal geometrical configuration of bodies, the symmetry of their physical properties, and the symmetry of physical fields. Physical phenomena or processes may involve all these factors, and it is therefore of interest to inquire what principles govern the relationships of their symmetry. The principles described here may appear self-evident, especially in simple cases, but they are basic concepts in physics, and it is thought worth while to state them explicitly, as an aid when physical notions are to be applied to subjects such as structural geology. Since symmetry has received the greatest attention in the study of crystals, the symmetry principles in physics have been based largely on experience in this field.

The most basic concept in physical crystallography is *Neumann's principle* (Voigt, 1928, p. 3; Love, 1927, p. 155; Nye, 1957, p. 20), which often is not stated explicitly but is taken for granted. This principle correlates the symmetry of the morphology of crystals and the symmetry of their physical properties. It may be stated as follows (Nye): "*The symmetry elements of any physical property of a crystal must include the symmetry elements of the point group of the crystal.*" Thus, the point group of the crystal class must be a subgroup common to the symmetry groups of all the physical properties of the crystal. This principle was based on the experience of the early experimenters (Voigt, 1928, p. 20). With the advent of X-ray crystal-structure analysis, it can now be understood as expressing the fact that the morphology and the physical properties both reflect the symmetry of the internal constitution of the crystals, the morphology being the most complete macroscopic expression of the particular symmetry in the structure. Neumann's principle does not forbid the physical property having symmetry elements lacked by the crystal. In fact, physical properties frequently have more symmetry than the point group of the crystal; for example, the optical refractive property of monoclinic and triclinic crystals (represented by the optical indicatrix) has orthorhombic symmetry (D_{2h}).

Important principles of symmetry governing physical phenomena were stated by Pierre Curie (1894). Curie stressed the value of applying the notions of symmetry, mainly developed in crystallography, to physical phenomena in general, demonstrating this especially with discussion of electric and magnetic fields and associated phenomena. Curie's principles have been elaborated by Jaeger (1920), while brief statements of some of them have been made elsewhere (for example, Bouasse, 1909; Nicolle, 1950, 1955). Recently, Shubnikov (1956) has also reviewed Curie's work in the field of symmetry, and other Russian workers have invoked Curie's principles of symmetry, especially in connection with piezoelectricity (for example, Zheludev, 1957a; Koptsik, 1957), while Shafranovskii (1956) has also drawn attention to their relevance in structural petrology.

It is convenient to discuss Curie's principles in three parts:

(1) Curie first introduces the notion that there is a certain "dissymmetry" that is characteristic of a physical phenomenon (Curie, 1894, p. 400):

> "The characteristic symmetry of a phenomenon is the maximum symmetry com-

patible with the existence of the phenomenon.

"A phenomenon can exist in a medium which possesses its characteristic symmetry or that of one of the subgroups of its characteristic symmetry.

"In other words, certain elements of symmetry can coexist with certain phenomena but they are not necessary. What is necessary is that certain elements of symmetry do not exist. It is the dissymmetry that creates the phenomenon."[7]

Pasteur (1861) introduced the term "dissymmetry" for the absence of symmetry elements of the second sort, but Curie uses it more generally to mean the absence of particular symmetry elements of either sort. A "physical phenomenon" here may be taken to be the appearance, in the medium, of a physical state which can be described in terms of a certain physical quantity, for example, an electric field, or an electric polarization, an elastic strain, or a flux of heat. Thus, in accordance with the above principle, an electric field (characterized by point group $C_{\infty v}$) cannot exist in a medium having a center of symmetry; any medium in which an electric field exists will lack a center of symmetry by virtue of its electrical polarization, if not otherwise. Similarly, a spontaneous electrical polarization can arise in a medium, the point symmetry of which is any subgroup of $C_{\infty v}$ (for example, tourmaline, C_{3v}), when it is heated.

2. Curie further states the following principle of superposition (p. 401):

"When several phenomena of different nature are superposed in the same system, the dissymmetries are added. Therefore, there only remain as symmetry elements in the system those which are common to each phenomenon taken separately."[8]

In other words, the resultant symmetry comprises those symmetry elements common to the several phenomena, forming a subgroup common to the symmetry groups possessed by the phenomena separately. Only similar elements occurring in the same orientation and position are to be considered common. For example, when a magnetic field (point symmetry $C_{\infty h}$) is superposed on an electric field (point symmetry $C_{\infty v}$) so that their directions (or infinite-fold axes) are at right angles, the resultant common symmetry C_{1h} consists of a single plane of symmetry, containing the direction of the electric field and being normal to the direction of the magnetic field. Similarly, a displacement field compounded of a rotation ($C_{\infty h}$) and a simple strain ($D_{\infty h}$) normal to the rotation axis has the monoclinic symmetry C_{2h}.

A similar superposition principle may be enunciated to take into account the symmetry of the medium itself. Although not stated explicitly by Curie, it is a corollary of his first principle, and its statement by Nye (1957, p. 245) for crystals may be generalized as follows: A medium under an external influence will exhibit only those symmetry elements that are common to the medium without the influence and to the influence without the medium. For example, if a crystal of halite is subjected to a simple compressive stress parallel to a three-fold axis, the resultant point symmetry of the deformed halite is D_{3d}, a subgroup common to both the symmetry of the unstressed halite (O_h) and the symmetry of the stress ($D_{\infty h}$).

3. Curie makes these statements about the conservation of symmetry in "causes" and "effects" (p. 401):

"When certain causes produce certain effects, the symmetry elements of the causes must be found again in the effects produced.

"When certain effects show a certain dissymmetry, this dissymmetry must also be found in the causes which have given rise to them.

"The converse of these two propositions is not true, at least in practice; that is to say, the effects produced can be more symmetrical than the causes."[9]

Jaeger (1920, p. 95) has expressed this as follows: ". . . *the effects may occasionally have the same or a higher symmetry than the causes, but the last cannot have a higher symmetry than the effects produced.*

"It is, moreover, worth remarking in this connection, that symmetry-properties which are present in *all* causes, and in *all* circumstances governing a certain phenomenon, are necessarily always found in the effects. However, we must always be sure that the number of causes consideered is really complete; evidently it is in many cases hardly possible to get full assurance of this."

For illustration, we may consider the follow-

[7] Literal translation from the French by M. S. Paterson

[8] Literal translation from the French by M. S. Paterson

[9] Literal translation from the French by M. S. Paterson

ing example. If a homogeneous triaxial stress, having the point symmetry D_{2h} of a general ellipsoid, is applied to an isotropic medium, it gives rise to a uniform strain which also has the point symmetry D_{2h} of a general ellipsoid; the absence of particular symmetry elements (such as infinite-fold axes) in the ellipsoid representing the strain necessarily implies the absence of such elements in the ellipsoid representing the stress. As a further example, the application of a stress to any body gives rise to a change in temperature which, being a scalar and having the point symmetry $K_{\infty h}$, is more symmetrical than the stress, again in accordance with the general rule above.

However, in general it may not be possible to distinguish between "causes" and "effects" among the quantities involved in a given physical system. Moreover, the symmetry of the medium must be taken into account. These difficulties may be resolved by considering the description of any physical situation to be the statement of a relationship or equation for the physical quantities concerned, in which parameters describing the relevant physical quantities of the medium also appear. Sometimes, it may be convenient to consider the independent variables in such an equation to be "causes" and the dependent variables to be "effects"; the symmetry of the "effects" then must include at least the elements that are common to the "causes" and the physical properties of the medium.

As Shubnikov (1956) has pointed out, this can be reduced to a single principle of superposition. The resultant symmetry in a given physical system is that which is common to the individual physical fields and properties that must be specified in a complete description of the system. Curie's statements above can then be understood by taking the symmetry of the "effects" to be the resultant symmetry of the physical system and the symmetries of the "causes" to be the individual symmetries of the various independent factors (physical properties or fields) that can be distinguished in any description of the system. For example, if we consider a crystal of halite in a certain state of elastic strain, the symmetry of the applied stress (or strain) and the symmetry of the elastic properties of the halite can be regarded as the symmetries of the "causes" (these being the independent factors contributing to the system), while the symmetry of the resulting strained (and stressed) halite is the symmetry of the "effects."

Thus, we may summarize Curie's symmetry principles as follows: *The symmetry of any physical system must include those symmetry elements that are common to all the independent factors (physical fields and physical properties of the medium) that contribute to the system, and it may include additional symmetry elements; however, any symmetry elements absent in the system must also be absent in at least one of the independent contributing factors.*[10]

PART II. APPLICATION OF SYMMETRY ARGUMENTS TO DEFORMED ROCKS

Introduction

Field and microscope observations indicate that certain bodies of rock have undergone marked permanent deformation while in the solid state. The characteristic features that define the fabric of such a body are described in detail. The basic problem in the study of a rock of this kind is to infer as far as possible from the observations on its fabric what its history of deformation has been. After considering the nature of the fabrics and of the quantities that characterize the deformation, we shall discuss how symmetry arguments can help to solve this problem. The symmetry considerations themselves do not establish that a rock has been deformed; this must first be decided on other grounds. The symmetry arguments only permit deductions from the fabric symmetry concerning the symmetry of the quantities characterizing the deformation (or whatever other factors are considered to have contributed to the emergence of the present features of the rock).

The mechanisms whereby rocks become permanently deformed are imperfectly understood, but many processes such as translation and twin gliding in the crystals, slip and rotation between grains or groups of grains, cataclasis, recrystallization, and neomineralization may be involved according to circumstances. All such processes whereby a body changes its configuration involve the relative movements of parts of the body, termed by Sander (1930, p. 115) *componental movements* (*Teilbewegungen*). Sander distinguishes two kinds of componental movements: (1) direct componental movements, in which the relative movements

[10] A more concise statement of this principle is given by Melvin (1951, p. 236–239), as follows: " 'Cause' symmetry is a subgroup in 'effect' symmetry." Other important symmetry principles are outlined by Melvin (p. 239–240).

are highly correlated in a direct way, as in gliding processes; (2) indirect componental movements, in which the correlation of relative movements can only be defined statistically, as in the ionic diffusion processes involved in recrystallization.

The relative importance of these various processes in the history of most deformed rocks is unknown, but it may be presumed to influence their patterns of preferred orientation of grains or other structural features. Thus the patterns of preferred orientation of grains that arise when diffusion processes predominate will probably be different from those to which deformation by translation and twin gliding lead. However, our concern here is not with the patterns of preferred orientations themselves but with their symmetry, and from Curie's principles we may expect the *symmetry* of patterns of preferred orientation arising in given circumstances to be the same whatever the mechanism of deformation.

The deformation of a body of rock may be effectively continuous on a large scale, but it will usually involve discontinuities or heterogeneities of movement on a small scale, corresponding, for example, to the relative movements of parts of grains in gliding or of groups of grains separated by shear surfaces. These discontinuities and heterogeneities of movement frequently leave an imprint in the rock in the form of heterogeneities of structure, which on a larger scale can be considered as pervasive or penetrative of the whole body of rock. Similarly, the reorientation of grains by gliding or diffusion processes within or between the grains during deformation may lead to a preferred orientation which is to some extent an imprint of the deformation. All such rocks, which bear evidence of having undergone appreciable permanent deformation, are called *tectonites*, a term introduced by Sander (1930, p. 62) for rocks "deren Teilbewegung im Gefüge summierbar ist." Various kinds of tectonites have been distinguished on descriptive or genetic bases (Sander, 1930, p. 58, 276; Knopf and Ingerson, 1938, p. 38–40), but in the present paper the term tectonite is used in a general way to refer to any rock whose internal geometrical features have resulted from appreciable permanent deformation by direct or indirect componental movements in the solid state. This restriction excludes weakly deformed sedimentary and igneous rocks with their special problems of analysis and interpretation; tectonites discussed here are almost exclusively metamorphic in origin (exceptions are halite and anhydrite rocks in salt domes and ice in glaciers).

The general problem is therefore to deduce as far as possible from the observed fabric the history of the deformation of a tectonite and the circumstances in which the deformation occurred. It is important to distinguish the various interrelated factors in terms of which the fabric of a particular body of rock may be interpreted, especially the following:

(1) The system of forces acting on the body of rock. These may be applied to the boundaries of the body of rock via adjacent bodies by some tectonic disturbance, or they may be gravitational forces acting on the mass of the rock itself.

(2) The stress field within the body of rock, which arises from the system of forces acting on it.

(3) The geometrical transformation expressing the displacement of each point within the body of rock during the deformation, from which local rotations, strains, etc., can be derived.

(4) The initial fabric (including the physical properties of the rock which determine its behavior under stress).

We shall be concerned mainly with the problem of deducing what geometrical transformation the rock must have undergone, given a certain initial fabric—that is, with the *kinematic interpretation* of tectonite fabrics. It is a more difficult problem to proceed to the further stages of deducing first the stress field in the rock during its deformation and then the system of forces connected with the force field. While the kinematic problem is one of geometry, the latter dynamic problems require a knowledge of appropriate physical properties of the rock, such as the parameters in the equations relating stress and strain during plastic deformation. In some cases where diffusion processes (indirect componental movements) predominate, it may be simpler to attempt the direct correlation of fabric and stress field.

Tectonite Fabrics

General. The term *fabric* (*Gefüge*) is used by Sander (1930, p. 1; 1948, p. 2) for both the geometrical and the physical spatial data of an aggregate. The concept of fabric includes therefore not only the form or geometrical arrangement of elemental parts (*gestaltliche Gefüge*) but also the functional or behavioral aspect

(*funktionale Gefüge*) which is concerned with the directional physical properties that follow from this geometrical organization of matter. The term fabric is accordingly used here to cover both aspects, but we shall be dealing mainly with geometrical features.

Statistical homogeneity on a given scale is usually implied in the concept of fabric, and therefore the determination of fabric involves observations on a homogeneous domain. Any tectonite can be divided into parts or *domains* that are statistically homogeneous with respect to their structure on a given scale.

In its geometrical aspect, the fabric of an aggregate is analogous to the structure of a crystal. While a crystal can be considered as a homogeneous continuum on most scales, understanding of its geometry and symmetry is possible only when the crystal is viewed more rigorously as a periodic array of structural discontinuities on the ionic or molecular scale. Similarly, structurally homogeneous bodies of rock must be examined in terms of their small-scale discontinuities or local heterogeneities before the concept of fabric becomes meaningful. A fabric, like a crystal structure, is a three-dimensional array of discontinuities or heterogeneities in structure. Statistical homogeneity of the body of rock means that this array is on the average everywhere the same, on the scale concerned.

The aspect of the fabric most conveniently studied geometrically is the *preferred orientation* of these small-scale features of the aggregate. There are several kinds of easily measurable small-scale geometrical features: (1) lattice planes and lines within individual grains, (2) shapes of inequant grains, (3) arrangements of grains of particular kinds into layers, linear streaks, and other inequant configurations.

Before such features define by their spatial array a homogeneous fabric, they must be *penetrative;* that is, *on the scale of the domain* under consideration, they must be repeated statistically at imperceptible distances so that they effectively pervade the body and are present in the same average orientation in every sample. If the body is sampled on a smaller scale, the same features may not show constant preferred orientation; on such a scale, the body can no longer be considered to be statistically homogeneous, and the features become nonpenetrative.

Fabric elements. The individual geometrical features of a fabric, which on the scale of a whole domain become penetrative, will be termed *fabric elements.* This term has been used by Sander (1948, p. 5) to denote the actual equivalent domains of which a body is composed, while Fairbairn (1949, p. 3) defines a fabric element as "a single crystal or group of crystals which act as a unit with respect to the forces applied to it." Neither definition embraces statistically pervasive features such as foliations and lineations since such features are not domains of a body but are surfaces and lines of structural discontinuity within or between domains of the body. All fabric elements can be viewed on some scale as surfaces and lines of structural discontinuity. For example, features defined by planar and linear preferred orientations of inequant grains are, in detail, also penetrative families of surfaces of discontinuity defined by the grain boundaries. Likewise, lattice planes and lines in individual crystals are planes and lines of discontinuity on the ionic or molecular scale.

In tectonite fabrics, we recognize two kinds of fabric elements:

(1) *Crystallographic fabric elements.* These are lattice planes or lines in individual grains, *e.g.*, {001} planes in mica (determined from visible cleavages), $\{01\bar{1}2\}$ planes in calcite and $\{02\bar{2}1\}$ planes in dolomite (determined from visible twin lamellae), [0001] directions in quartz (determined as the optic axes), and [0001] directions in calcite and dolomite (determined as the optic axes or from known angular relations to cleavages and twin lamellae). These crystallographic planes and lines are to be thought of as pervading the whole grain and should not be confused with parallel individual visible discontinuities (cleavages, twin lamellae, etc.) that might be used in the practical determination of their attitude in a grain. On the other hand, twin lamellae, for instance, may also be treated as fabric elements in their own right.

(2) *Noncrystallographic fabric elements.* These are visible structural discontinuities or heterogeneities in an aggregate. Planar noncrystallographic elements are structures such as bedding and foliation (metamorphically produced surfaces of discontinuity in tectonites, for instance cleavage and schistosity), which are defined by preferred orientation of grain boundaries or by lithologic layering. Some aggregates contain more than one foliation. One of these may be parallel to axial planes of folds in another. Axial planes of folds may themselves become fabric elements on a larger scale.

Linear noncrystallographic fabric elements

are structures such as lineations and fold axes. Lineations can be defined by such features as lines of intersection of two planar fabric elements, crenulations in a planar fabric element, alignment of boundaries of elongated grains, and so on.

Graphic representation of fabrics. The attitude of any fabric element can be specified by its angular relation to geographic axes or to chosen orthogonal axes, called *fabric axes*, generally labeled *a*, *b*, and *c*. The measurements that specify the attitudes of fabric elements are termed *fabric data*. These are usually plotted in a Schmidt "equal-area" projection[11] in a way similar to that in which the geometrical properties of crystals are plotted in a stereographic projection, except that the lower hemisphere instead of the upper is used for the fabric data.

The use of the equal-area projection facilitates the contouring of the density distribution of the points representing the data, which is necessary for the statistical definition of the fabric. Such a projection of the fabric divorces orientation from location so that, as in the stereographic projection of crystal data, the point symmetry is more readily discerned. If the fabric is centrosymmetric, only one hemisphere need be used in the projection.

The symmetry of the preferred orientation of the fabric elements is determined by the symmetry elements that can be observed in such a projection. In this connection, only the form of the main features of the patterns (such as well-defined girdles or strong maxima) should be considered, since the diagrams usually lack all symmetry if every detail is taken into account. The main features are reproducible

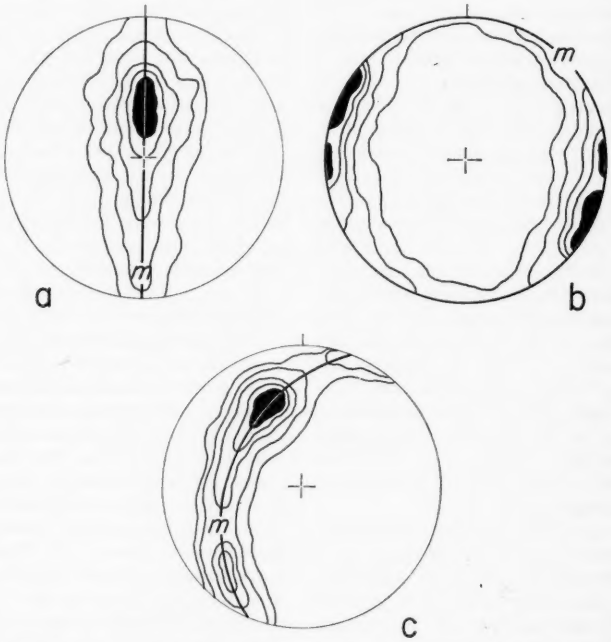

Figure 5. Examples of planes of symmetry (*m*) in contoured equal-area projections (a) normal to plane of projection, (b) parallel to plane of projection, (c) oblique to plane of projection

[11] For details of plotting and contouring fabric data, the reader is referred to Knopf and Ingerson (1938, p. 226–262) and Fairbairn (1949, p. 278–291); and for a discussion of fabric axes to Turner and Verhoogen (1960, p. 620).

in different sets of adequate data from the same homogeneous body, whereas the detail of the contour patterns is not. Examples of diagrams showing planes of symmetry in three orientations relative to the plane of projection (normal, parallel, and oblique, respectively) are given in Figures 5*a*, *b*, *c*.

Subfabrics. A fabric is defined by the spatial array of all its elements. However, it is frequently sufficient or convenient to consider the array of only one kind of element. This array is called a *subfabric.*[12] A subfrabric may have the same symmetry as the whole, or it may have higher symmetry. This is in some respects analogous to the crystallographic situation in which a given lattice plane may have the same symmetry as the crystal or a higher symmetry, depending on whether it corresponds to a general or a special form (*e.g.*, {111} and {110}, respectively, in the crystal class 222).

Many kinds of element can generally be distinguished in a fabric. In practice, a few easily measurable elements are chosen, and the subfabrics defined by these are combined to define the geometrical properties of the complete fabric. Fabrics in which all subfabrics agree in symmetry are termed by Sander (1930, p. 165) *homotactic fabrics;* those in which the subfabrics do not agree are termed *heterotactic fabrics.*

Statistical Symmetry of Tectonite Fabrics

General. The symmetry of a fabric is a space symmetry although it can be specified in terms of point symmetry since restriction to statistical symmetry implies only that arbitrary translations in all directions are added to the point symmetry operations. In heterogeneous situations such translations are absent, and point symmetry becomes relevant. Certain heterogeneous portions of a fabric (for instance, an individual small fold in a large body similarly folded throughout) may show point symmetry isomorphous with the space symmetry of the fabric of which the heterogeneous portion is a small part. Likewise individual fabric elements have point group symmetry, but this clearly can be different from the space group symmetry of their three-dimensional array in a subfabric.

Symmetry of fabric elements. Before considering the symmetry of the fabric, we shall discuss the symmetry of the fabric elements themselves, since this may have a bearing on the symmetry of their array. Noncrystallographic fabric elements such as foliations, lineations, fold aces, axial planes, etc., always have symmetry $D_{\infty h}$—the symmetry of a nonpolar line. Bedding is one of the few polar fabric elements encountered in tectonites (others are flow structure inherited from sedimentary fabrics, and slickensides), and this is generally polar only in a stratigraphic sense where top and bottom of the beds are known; on the other hand, a graded bed is a truly polar feature with symmetry $C_{\infty v}$. However, the polar nature of such fabric elements has generally been ignored up to date, and they have been treated as if they were nonpolar.

Certain combinations of related noncrystallographic fabric elements can have lower symmetry than $D_{\infty h}$. For instance, the full specification of a fold includes an axis and an axial plane in which the axis lies. In projection, this combination has symmetry D_{2h} (Fig. 6*a*). However, in practice it is better to treat these characteristics of a fold as separate fabric elements because the subfabric defined by the array of axes can have symmetry different from that defined by the array of axial planes measured in the same folds. Diagrammatic examples of three possible combinations of axes with axial planes are given, with corresponding symmetry, in Figures 6*b–d*.

Crystallographic fabric elements can have the symmetry $D_{\infty h}$ of a nonpolar line, *e.g.*, {001} planes of mica (represented by their normals) and [0001] axes in quartz. Other crystallographic fabric elements, however, may have other symmetries. Thus, the three $\{01\bar{1}2\}$ planes, parallel to visible twin lamellae, in calcite define a fabric element with symmetry D_{3d}. Again, the {110} planes of amphibole constitute a fabric element with symmetry D_{2h}. Here it is important to distinguish whether one takes the fabric element to be an individual plane or line, or the combination of equivalent planes or lines that occurs in the single crystal. In the former case the symmetry, as for one $(01\bar{1}2)$ plane of calcite, is still $D_{\infty h}$.

Where fabric elements with crystallographic symmetry have a preferred orientation, the subfabric that they define can have pseudocrystallographic point symmetry, which is unlikely to arise from noncrystallographic elements. However, pseudocrystallographic symmetry does not necessarily follow from the preferred orientation of such fabric elements;

[12] We propose this term as a translation of Sander's term *Teilgefüge*, instead of *partial fabric.*

for example, a calcite aggregate in which an *a* axis for each grain had for some reason become aligned parallel to a unique direction would have symmetry $D_{\infty h}$ if the *c* axes had not been aligned. Pseudocrystalline symmetry follows if there is preferred orientation of a set of equivalent nonunique crystallographic planes or lines.

An example of a tectonite with pseudocrystallographic symmetry is the Poughquag quartzite studied by Higgs, Friedman, and Gebhart (1960, Fig. 6E, p. 285). In this rock, the quartz [0001] axes, $\{11\bar{2}0\}$ planes, and $\{10\bar{1}1\}$ planes have preferred orientations which respectively define subfabrics with symmetry $D_{\infty h}$, D_{6h} and D_{3d}. The fabric made up of the combination of these subfabrics has symmetry D_{3d}, which is still higher than the symmetry, D_3, of quartz itself. It is possible to determine whether the fabric of the rock is really D_3 or D_{3d} only by using techniques, such as the measurement of piezoelectric properties, that can distinguish polar directions. If such a rock is found to be piezoelectric, the symmetry of its fabric is most likely to be isomorphous with the point group D_{3h} owing to the presence of left-handed and right-handed

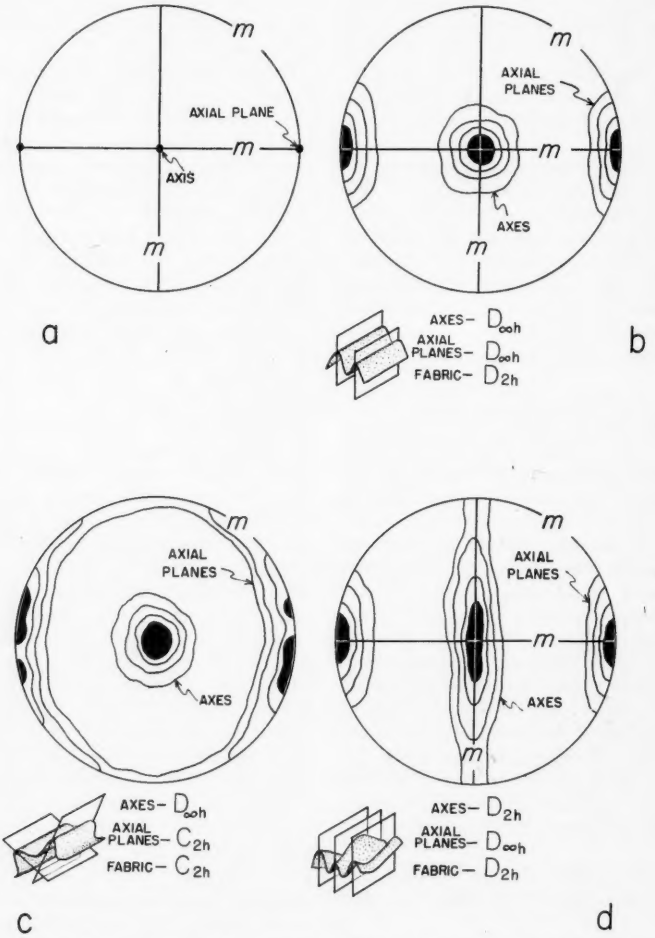

Figure 6. Symmetry of different combinations of axes and axial planes of folds (m = planes of symmetry)

quartz grains in equal numbers (Parkhomenko, 1959).

A similar example for calcite is given by the experimentally deformed Yule marble described by Turner, Griggs, Clark, and Dixon (1956, Fig. 11A, p. 1277). Here the crystallographic *a* and *c* axes are preferentially oriented to define symmetry D_{6h} for the combination of their subfabrics. Again, this symmetry is higher than the crystallographic symmetry of calcite (D_{3d}), but it is possible that by taking into account other subfabrics (*e.g.*, of $\{01\bar{1}2\}$ planes) a fabric might be defined with the symmetry of calcite.

Symmetry of subfabrics. In principle, subfabrics may have any of the symmetry groups of a homogeneous continuum, as discussed above. However, in general practice, so far, only centrosymmetric fabric elements have been measured, and the possible arrays of these therefore constitute the centrosymmetric symmetry classes for fabrics listed by Weissenberg. The study of natural tectonites has shown that with respect to fabric elements of the kinds

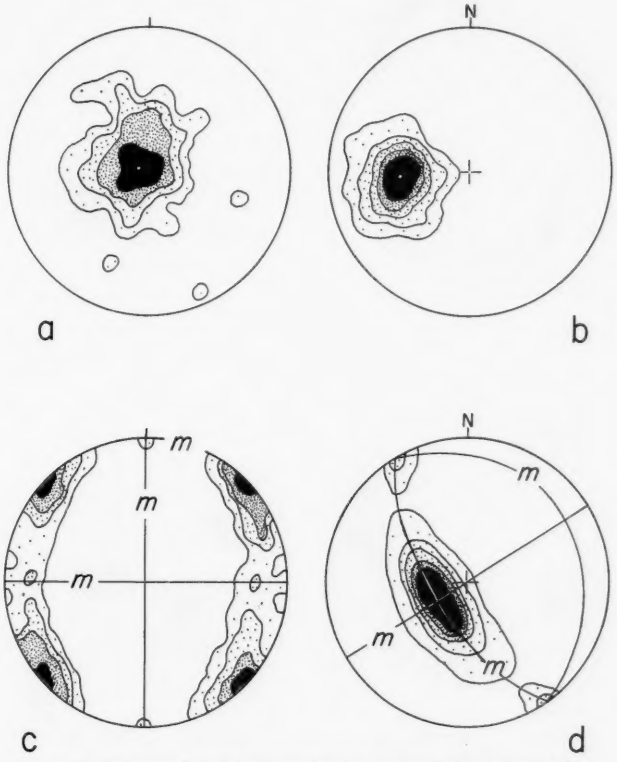

Figure 7. Contoured equal-area projections illustrating different kinds of symmetry observed in subfabrics (*m* = planes of symmetry)

Axial subfabrics ($D_{\infty h}$): (a) 100 [e_1:e_2] axes of calcite in marble from Barstow, California; contours 1–3–5–7% per 1% area (Weiss, 1954, p. 33), (b) 171 lineations in quartzite from Ballachulish, Scottish Highlands; contours 1–4–7–10–13% per 1% area

Orthorhombic subfabrics (D_{2h}): (c) 300 [0001] axes of quartz in mylonitized quartzite in the Moine thrust zone, Stock of Glencoul, Scottish Highlands; contours 1–3–5–8% per 1% area (measurements by J. M. Christie), (d) 1000 poles to foliation in gneiss at Turoka, Kenya; contours 1–3–5–7–9% per 1% area (Weiss, 1954, p. 46)

discussed in the previous sections the natural tectonite fabrics have very few different kinds of symmetry. Excluding pseudocrystallographic symmetries defined by fabric elements with other than axial symmetry ($D_{\infty h}$), only 1-, 2- and ∞-fold rotation axes and planes and centers of symmetry can be expected to arise. (*Cf.* Sander, 1950, p. 26.) Therefore, by putting $n = 1$, 2, and ∞ in Weissenberg's list of centrosymmetric groups, we obtain only $S_2(C_i)$, C_{2h}, D_{2h}, $D_{\infty h}$, and $K_{\infty h}$.

These five symmetry classes are the observed classes of subfabrics of tectonites and are named as follows (*see also* Sander, 1930, p. 146):

1. *Spherical symmetry* $K_{\infty h}$: such fabrics have the symmetry of a sphere. Ideally this symmetry is shown by a random orientation of fabric elements and is therefore generally only approached in natural tectonites. Some hornfelses have subfabrics approaching this symmetry.

2. *Axial symmetry* $D_{\infty h}$: such fabrics have

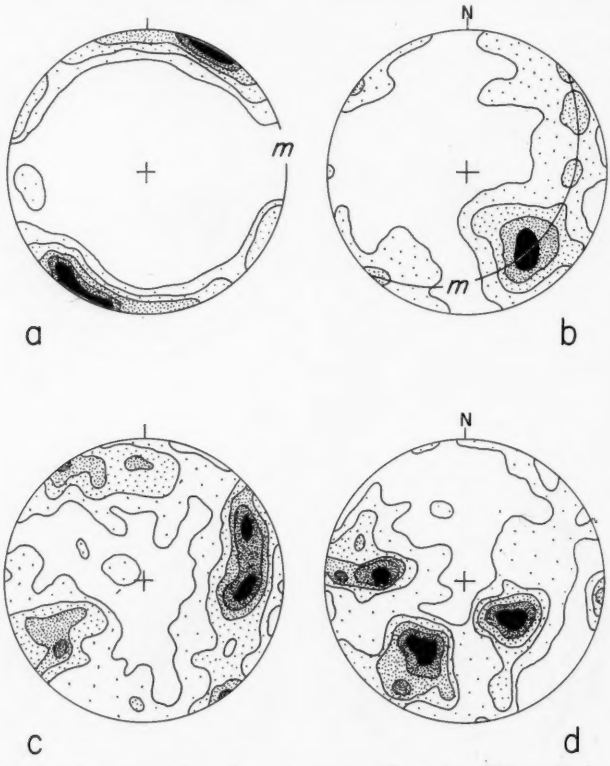

Figure 8. Contoured equal-area projections illustrating different kinds of symmetry observed in subfabrics (m = planes of symmetry)

Monoclinic subfabrics (C_{2h}): (a) 255 [0001] axes of quartz in mylonitized quartzite from Barstow, California; contours 1–3–5–7–9% per 1% area (Weiss, 1959, p. 51), (b) 2000 poles to foliation in schists and quartzites, Loch Leven, Scottish Highlands; contours 1–2–3–4% per 1% area (Weiss and McIntyre, 1957, p. 588)

Triclinic subfabrics ($S_2 = C_i$): (c) 416 [0001] axes of quartz in stretched quartzite pebble from Panamint Range, California; contains 1–2–3–4–5% per 1% area, (d) 193 poles to foliation in gneisses and schists near Lake O'Keefe, East Central Quebec, Canada; contours 1–2–3–4–5% per 1% area (measurements by G. Gastil and L. E. Weiss)

the symmetry of a cylinder; that is, in projection there is a unique axis that is the line of intersection of an infinite number of planes of symmetry and is normal to another plane of symmetry. Examples of subfabrics with this symmetry are shown in Figures 7*a* and *b*.

3. *Orthorhombic symmetry* D_{2h}: in projection such fabrics have three mutually perpendicular

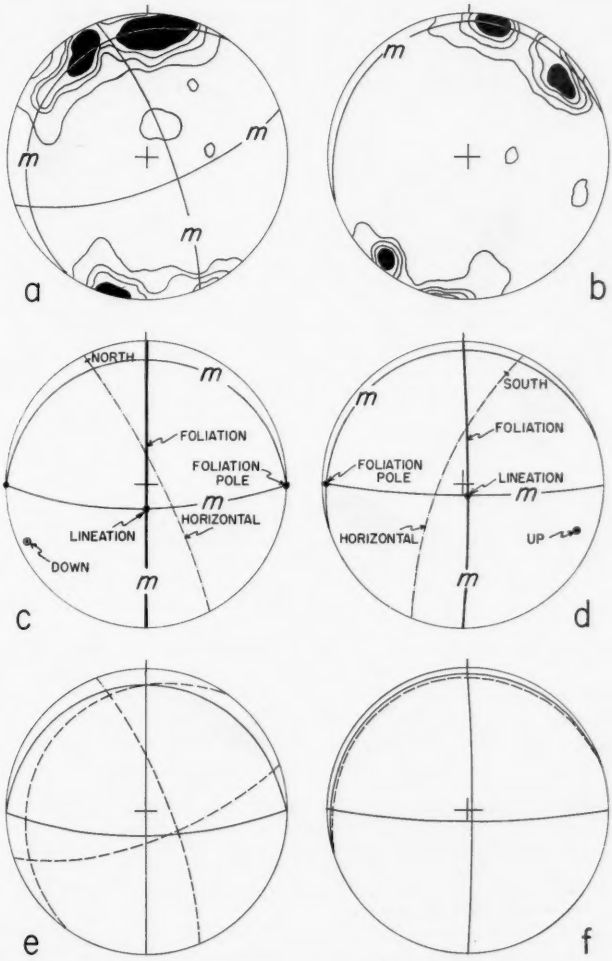

Figure 9. Data from two quartz tectonites from Turoka, Kenya (m = planes of symmetry). (a) Subfabric of 250 [0001] axes of quartz from quartzites; contours 1–3–5–7% per 1% area: orthorhombic pattern, (b) Subfabric of 300 [0001] axes of quartz from quartzite; contours 1–5–9–13% per 1% area: monoclinic pattern, (c) Foliation pole and lineation from specimen in a: orthorhombic pattern, (d) Foliation pole and lineation from specimen in b: orthorhombic pattern, (e) Superposition of planes of symmetry from a and c: none in common; fabric triclinic, (f) Superposition of planes of symmetry from b and d; one in common; fabric monoclinic

planes of symmetry. Examples of subfabrics with this symmetry are shown in Figures 7*c* and *d*.

4. *Monoclinic symmetry* C_{2h}: in projection, such fabrics have a single plane of symmetry, as shown in examples in Figures 8*a* and *b*.

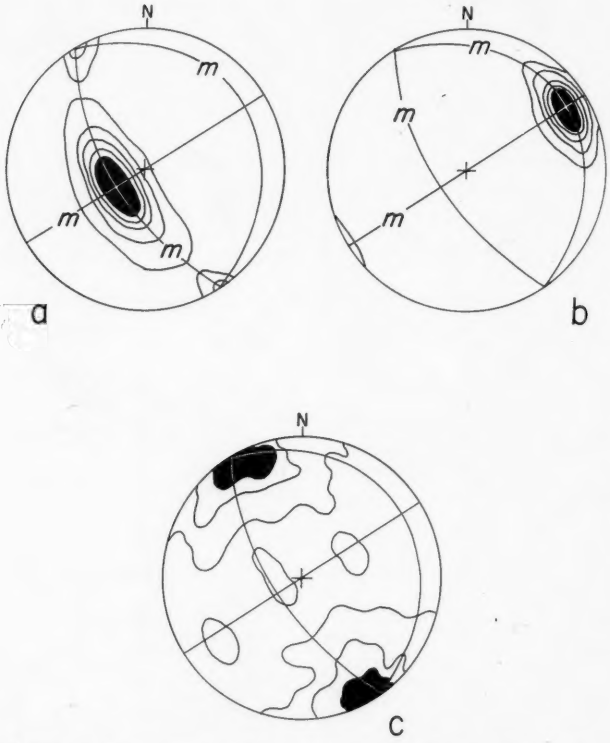

Figure 10. Data from schists and gneisses near Turoka, Kenya (m = planes of symmetry). (a) 1000 poles to foliation; contours 1–3–5–7–9% per 1% area: orthorhombic pattern, (b) 800 lineations; contours 1–5–10–15–20% per 1% area: orthorhombic pattern, (c) 2800 [0001] axes of quartz (synoptic); contours 3–6–9% per 1% area: axial (?) or orthorhombic pattern

5. *Triclinic symmetry* $S_2 = C_i$: no planes of symmetry, as shown in examples in Figures 8*c* and *d*.

Except for the first two, these symmetry types are named from the holohedral crystal classes of the same symmetry, viz., m m m, 2/m and $\bar{1}$, respectively.

Symmetry of fabrics. The symmetry of the fabric is given by the symmetry elements that are common when we superimpose the symmetries of the subfabrics for the various kinds of fabric elements being taken into account. That is, similarly oriented symmetry elements that are common to all the subfabrics are also symmetry elements of the fabric. Symmetry elements not present in *all* subfabrics are not symmetry elements of the fabric, and so a fabric cannot have symmetry higher than that of any of its subfabrics. Therefore, the five important types of symmetry listed for subfabrics above will again be the important types that occur for fabrics.

Experience shows that very few of the possible subfabrics of a given fabric need be

considered in order to determine the symmetry of the fabric. Thus if account is taken of the preferred orientation of axial crystallographic elements of two major, dissimilar minerals (*e.g.*, quartz and mica or calcite and mica) together with that of noncrystallographic elements such as foliations and lineations, a total symmetry is generally defined which remains unchanged if other subfabrics are taken into account.

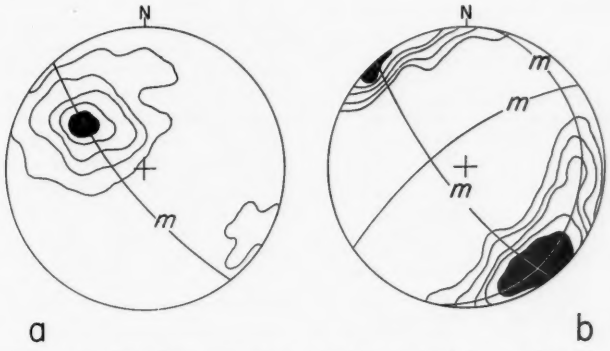

Figure 11. Data from quartzites and schists at Loch Leven, Scottish Highlands (m = planes of symmetry). (a) 500 lineations (early); contours 1–3–5–7–9% per 1% area: axial pattern, (b) 700 poles "fracture cleavage" (late); contours 1–2–3–4–5% per 1% area: orthorhombic pattern

Examples. The following are examples of the fabrics of natural tectonites, analyzed on various scales in terms of subfabrics:

1. *Tectonites from Turoka, Kenya* (Weiss, 1959). The rocks are foliated and lineated quartzite in gneiss and schist. The data from two hand specimens are shown in Figure 9. Figures 9*a* and *b* show preferred orientation of [0001] axes of quartz. The first of these subfabrics (Fig. 9*a*) is broadly orthorhombic, and three planes of symmetry can be drawn, as shown (m). The second has three maxima and is monoclinic with a single plane of symmetry (m in Fig. 9*b*). The planes of symmetry for the combined subfabrics of foliation and lineation are shown for the two specimens, respectively, in Figures 9*c* and *d* (geographical orientation of plane of projection is also given). These are orthorhombic with the planes of symmetry shown. The planes of symmetry from Figure 9*a* (broken lines) and Figure 9*c* (full lines) are combined in Figure 9*e* for the first specimen. None is in common, so that the total fabric lacks planes of symmetry and is triclinic. The planes of symmetry from Figure 9*b* (broken line) and Figure 9*d* (full lines) are combined in Figure 9*f* for the second specimen. One plane of symmetry is common to both subfabrics. This is also a plane of symmetry for the fabric defined by the three subfabrics, which is therefore monoclinic. For either specimen, inclusion of other subfabrics (for instance, preferred orientation of {001} planes of mica) could further lower the symmetry of the total fabric.

In Figure 10, data from rocks in the same region collected over an area of 50 square miles are projected on a horizontal plane. Figure 10*a* shows the poles to 1000 foliation surfaces, and figure 10*b* shows 800 lineations. Both diagrams are clearly orthorhombic with planes of symmetry identically situated in each diagram, as shown (m). Figure 10*c* is a synoptic diagram of 2800 [0001] axes of quartz from 12 specimens scattered throughout the area. The diagram is broadly axial with an ∞-fold axis parallel to the single strong maximum. Superposition of the planes of symmetry from Figures 10*a* and *b* shows that these are also planes of symmetry for the fabric of the large domain. Thus, bodies of rock may have triclinic or monoclinic fabrics with respect to certain fabric elements (in these examples, foliation and lineation) in small domains (Figs. 9*e* and *f*), but be orthorhombic on a larger scale (Fig. 10*c*) where the preferred orientations of the same fabric elements are

concerned. Symmetry of fabric determined on one scale in a deformed body, therefore, should not be extrapolated to a larger scale.

2. *Tectonites from Loch Leven, Scottish Highlands* (Weiss and McIntyre, 1957). In Figure 11*a* are shown 500 early lineations measured in an area of 15 square miles of mica schist and quartzite. The diagram is effectively of axial symmetry, with a single strong maximum as the ∞-fold axis. "Fracture cleavage" (a later transgressive foliation) in the same rocks gives the pole diagram (700 poles) shown in Figure 11*b*. This diagram is clearly orthorhombic with three planes of symmetry (*m*) as shown. One of these planes of symmetry passes through the maximum of Figure 11*a* and is a plane of symmetry for the monoclinic fabric defined by these two subfabrics.

For additional discussion of symmetry of observed fabrics the reader is referred to Knopf and Ingerson (1938, p. 42–62).

Table 5. Noncrystallographic Elements of Initial Fabrics

Rock type	Fabric elements	
	Planar	Linear
Sedimentary	Bedding Cross-bedding Planar dimensional preferred orientation of grains	Ripple marks Groove casts Flute casts, etc. Intersections of normal and cross-bedding Linear dimensional preferred orientation of grains
Igneous	Bedding or compositional layering Flow layering Planar dimensional preferred orientation of grains	Flow lineation Linear dimensional preferred orientation of grains
Metamorphic (including tectonites)	Foliation (cleavage, schistosity, etc.) Axial planes of folds Planar dimensional preferred orientation of grains Inherited earlier planar structures	Lineation Fold axes Other linear structures (boudins, mullions, elongated pebbles, fossils, etc.) Inherited earlier linear structure

Factors in the Evolution of Tectonite Fabrics

General. In structural analysis, the tectonite fabric is viewed as the record or imprint of the rock's geological history, and it is the task of the geologist to extract from this record as much information as possible about the principal factors in the evolution of the rock—namely, the initial fabric, the pattern of the geometrical changes or deformations of the rock, and the influences that have brought about these changes. In the following sections we consider the nature of these factors and especially their symmetry.

Initial fabrics. Any type of rock can be deformed to become a tectonite, so the initial fabric may have any symmetry found in the fabrics of sedimentary, igneous, or metamorphic rocks. The most important noncrystallographic fabric elements found in these various rock types are listed in Table 5. In fabrics defined by such elements there may be, in addition, subfabrics defined by crystallographic fabric elements—that is, preferred orientations of lattice planes or lines in the grains.

The common types of symmetry to be expected in typical sedimentary and igneous rock fabrics are summarized in Table 6. A metamorphic rock that has already undergone a previous deformation can have any of the tectonite fabric symmetries discussed; other metamorphic rocks may be classed with igneous rocks with respect to symmetry.

In the interpretation of the final tectonite fabric, it is important to distinguish to what extent the symmetry of the initial physical property follows the initial geometrical fabric symmetry and, in particular, whether the geometrical discontinuities or heterogeneities that constitute the initial fabric correspond to heterogeneities in physical properties. In the former case, the fabric features will tend to be correlated with the local heterogeneities in the geometrical transformation that contribute to the "movement picture" discussed below; in the latter case, the initial geometrical fabric features will be mechanically ineffective and act simply as markers in the geometrical transformation.

Kinematics of deformation—the movement picture. In this section we shall be concerned with the purely geometrical or kinematic problem of describing the geometrical transformation that we have so far loosely called the "deformation." This consists basically of specifying the transition from an initial set of positions for the material points of the body to a final set of positions—that is, of specifying the *displacements* (Fig. 12). In geological

writing, the term "movement" can often be taken to refer to these displacements. However, the various features of the final rock fabric are not directly related to the displacements themselves, but to quantities such as strain, which can be derived from the displacements. Thus, the displacements themselves are of less importance than the differential displacements of points, as Sander sought to show by emphasizing the "componental movements" (*Teilbewegungen*).

ential displacement of points spaced equally in a particular direction is everywhere the same. However, we must consider transformations that are only statistically homogeneous and attempt to take into account also the local departures from homogeneity; the latter are essential since many geometrical features of a tectonite fabric are the imprint of local inhomogeneities in the displacement field.

The displacement of any point can be represented by a vector **u** (Fig. 12); the position of

Table 6. Symmetry of Initial Fabric

Symmetry	Examples	
	Sedimentary rocks	Igneous rocks
Spherical $K_{\infty h}$	Aggregates of equant grains, *e.g.*, massive sandstones with spherical quartz grains lacking preferred orientation of crystallographic directions	Aggregates with randomly oriented grains, *e.g.*, some granites and basalts (a hornfels is a metamorphic rock that can have this symmetry)
Axial $D_{\infty h}$	Bedded sediments of equant and inequant grains, *e.g.*, shales and limestones	Aggregates with bedding or nonlinear flow layering, or purely linear flow structures, *e.g.*, some lava flows, minor and major intrusive rocks
Orthorhombic D_{2h}	Rare	Rare
Monoclinic C_{2h}	Bedded sediments containing linear structures, *e.g.*, graywackes with groove and flute casts, imbrication, grain lineation	Aggregates containing laminar and linear flow structures, *e.g.*, some plutonic and volcanic rocks
Triclinic $S_2=C_i$	Bedded sediments containing nonparallel linear flow structures, convolute bedding, and so on	Aggregates containing irregular flow structures, *e.g.*, convolutions

If a displacement field is such that all points that fell on given straight lines or in given planes are again on straight lines or planes, the transformation is said to be affine or homogeneous. In such a transformation, the differ-

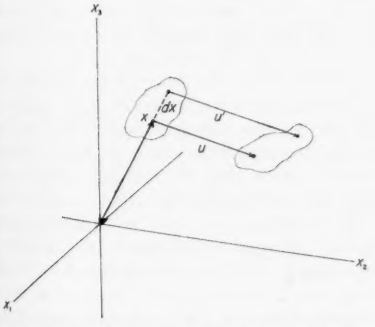

Figure 12. Displacement of a pair of points

the point itself relative to a given fixed frame of reference can also be represented by a vector **x** which joins the origin to the point. Then any point near **x** will undergo a displacement

$$\mathbf{u}'=\mathbf{u}+(\textit{Gradient}\ \mathbf{u})\ \mathbf{dx}$$

where **dx** is the vector representing the position of this point relative to the point **x**. Gradient **u** represents the rate of change of displacement **u** with change of position[13] and is a tensor quantity (having in general nine components instead of the three components of a vector). In an affine transformation, Gradient **u** is everywhere the same, and it is possible to write

[13] This analysis is meaningful only when **u** is a continuous and differentiable function of position; however, this is, in general, not a serious restriction since many apparent discontinuities may be considered as zones of rapid change. Gradient **u** is used here to denote the "vector gradient" of **u** (*see* Joos, 1934, p. 32), which is not to be confused with the gradient of a scalar.

it as the sum of two quantities, one representing a rotation and one a pure strain. This is a well-known result in the theory of finite homogeneous strain (Love, 1927, p. 59–73; Planck, 1932, p. 5–18). Thus, any homogeneous displacement field in a body can be analyzed into three parts: (1) a translation of the body as a whole; (2) a rotation of the body as a whole (a rigid-body rotation or, in geological terminology, an external rotation); (3) A pure strain, which can be represented by a strain ellipsoid and which in general will be a finite strain. The translational part need not be considered further since it is equivalent only to a shift of origin of coordinates and can have no influence on the fabric of the rock. We are interested only in the *relative displacement* of points in the body, that is, in

$$\mathbf{du} = \mathbf{u} - \mathbf{u}' = (\textit{Gradient}\ \mathbf{u})\ \mathbf{dx}$$

or, if the points are not close together, in

$$\Delta\mathbf{u} = \int_{\mathbf{x_2}}^{\mathbf{x_1}} (\textit{Gradient}\ \mathbf{u})\ \mathbf{dx},$$[14]

where $\mathbf{x_1}$ and $\mathbf{x_2}$ define the positions of the points or parts of the body, the relative displacement of which is in question. Therefore, $\Delta\mathbf{u}$ is an expression of Sander's *Teilbewegung* (componental movement), and, if we take the *Bewegungsbild* or (movement picture) to be the field of the componental movements, it follows logically that the "movement picture" is expressed by the field of $\Delta\mathbf{u}$. However, $\Delta\mathbf{u}$ is not a simple vector in three dimensions but is a vector in six dimensions—that is, it is a function of six variables, viz., the components of $\mathbf{x_1}$ and $\mathbf{x_2}$.

For many purposes, especially in correlating the deformation with the fabric, it is convenient to confine attention to Gradient **u**. In common usage, the "movement picture" seems to correspond to Gradient **u**. It will be seen below that it is the symmetry of Gradient **u**, or at least of part of it, that is relevant in determining the fabric symmetry.

In a strictly homogeneous transformation, Gradient **u** is equivalent to the combination of a rigid-body rotation and a pure strain. The rotation of the body as a whole can have no influence on the fabric, so that any aspects of the fabric related to a homogeneous transformation must therefore be related to the pure strain. As can be seen from the representation of strain as a strain ellipsoid, a pure homogeneous strain can have only spherical ($K_{\infty h}$), axial ($D_{\infty h}$), or orthorhombic (D_{2h}) point symmetry. Therefore, on symmetry arguments alone, the occurrence of monoclinic or triclinic fabrics cannot be ascribed to the influence of a strictly homogeneous transformation only. It is not sufficient to point to the monoclinic or triclinic symmetry that arises from the combination of the pure strain and the rotation, since the rigid-body rotation is irrelevant for the fabric. This anomalous situation is resolved by taking into account the local inhomogeneities in a statistically homogeneous field.

In a statistically homogeneous transformation, the displacement **u** at any point can be written as the sum of two vectors: (1) the displacement **U** in the strictly homogeneous field to which the actual field statistically approximates, and (2), the minor correction or perturbation $\delta\mathbf{U}$ that must be added to **U** to give the actual displacement at the point, *i.e.*, $\mathbf{u} = \mathbf{U} + \delta\mathbf{U}$. Likewise, the gradient of the displacement can be written as the sum of two parts, Gradient **U**+Gradient $\delta\mathbf{U}$. Since Gradient **U** represents a strictly homogeneous transformation, it can be resolved into a rigid-body rotation and a pure homogeneous strain, and as before we can disregard the rotation in interpreting a fabric. However, Gradient $\delta\mathbf{U}$ does not represent a homogeneous transformation; it varies from point to point and cannot be resolved into a rotation of the body as a whole and a homogeneous strain. Only in the neighborhood of a given point can Gradient $\delta\mathbf{U}$ be regarded as sensibly constant, so that it can be resolved into a pure rotation and a pure strain for the neighborhood of the point.[15] Thus, Gradient $\delta\mathbf{U}$ can be regarded as representing a local rotation and strain, the values of which vary from point to point and average out to zero over the whole body.

By neglecting the over-all translation and rotation of the body, we are in effect referring the deformation (now represented by Gradient $\delta\mathbf{U}$ plus the pure strain part of Gradient **U**) to internal rather than to external co-ordinates. Since the preferred orientation of fabric elements is defined only by their relation to one another, that is, by reference to internal co-ordinates, it is this aspect of the deformation

[14] It is assumed without proof that this integral is independent of path, which may not be valid where successive deformations are being considered.

[15] There is also a local translation associated with the $\delta\mathbf{U}$, but this can usually be disregarded.

that is especially relevant to fabric. The principal axes of the mean strain ellipsoid may be chosen as suitable internal orthogonal coordinate axes since they are parallel to lines of material particles which, statistically, were also orthogonal lines before the deformation. In general, all other lines and planes of material particles rotate relative to these axes (the "internal rotation" of Sander, 1948, p. 36).

As an example, let us consider a model of simple shearing in a single set of discrete parallel shear zones (Fig. 13). On the large scale, the transformation may appear to be strictly homogeneous and no fabric is discernible on this scale (Fig. 13*a*); any initial circle is transformed into an ellipse, as shown. However, examination on a finer scale (Figs. 13*b* and *c*) shows that the transformation is only statistically homogeneous. Figure 13*c* shows how the deformation may be analyzed as discussed above; the circles and ellipses (with corresponding rotations) on the left represent the actual Gradient **u**, which is equivalent to an ellipse representing Gradient **U** (everywhere the same) plus another ellipse representing Gradient **δU**, the local departures from homogeneity. if such a deformation took place in a rock, it is likely that fabric features would be developed that would reveal the local inhomogeneities of deformation (especially the shearing zones) and possibly also the over-all homogeneous strain. However, it is clear that the over-all rotation

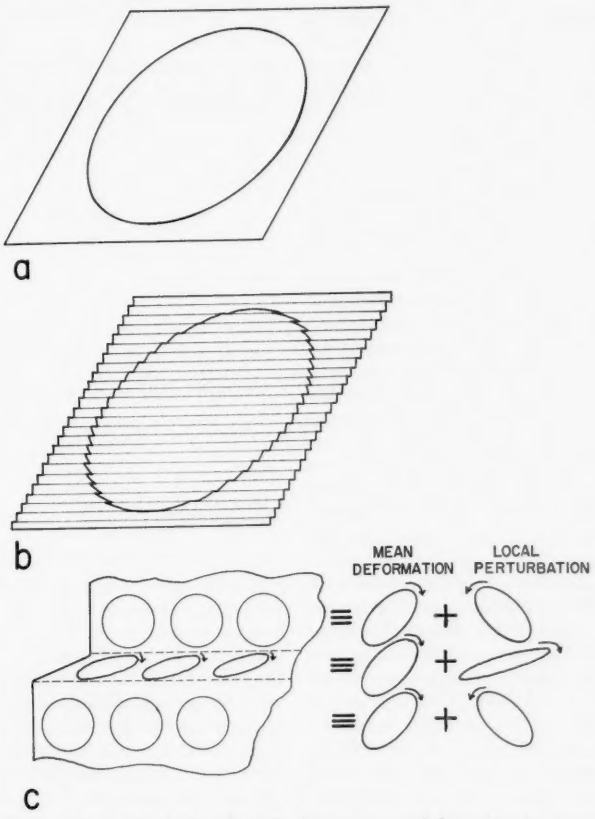

Figure 13. Analysis of statistically homogeneous deformation in terms of local heterogeneities

will leave no imprint in the fabric; for example, the fabric is the same whether or not Figure 13 is rotated through any chosen angle. It is only by referring to an external frame of reference (such as geographic axes) that the over-all rotation of a body can be detected.

The example shown in Figure 13 corresponds to plane deformation where no extension is assumed parallel to the principal axis of the ellipsoids normal to the plane of the figures. Such plane strains are rare in natural aggregates, but the principles just stated apply equally well to all cases of nonplane deformation.

It is of interest at this point to discuss the autocorrelation of the displacements—that is, the degree of consistency in the relative displacements $\Delta\mathbf{u}$ of pairs of points at a given spacing when all such pairs of points in the domain are considered. Again it is convenient to consider separately the displacements $\mathbf{U}$ corresponding to the strictly homogeneous part and the displacements $\delta\mathbf{U}$ corresponding to the local deviations from strict homogeneity in the actual statistically homogeneous transformation. The relative displacements $\Delta\mathbf{U}$ of pairs of points is the same in all parts of the domain, no matter what spacing of points is being considered. However, the relative displacements $\Delta(\delta\mathbf{U})$ are consistent only for closely spaced pairs of points within small regions of the domain; the mean value of $\Delta(\delta\mathbf{U})$ is zero when the average is taken over the whole domain or when widely spaced pairs of points are considered. That is, we can say that the displacements $\mathbf{U}$ are perfectly correlated over the whole domain but that the deviations $\delta\mathbf{U}$ are well correlated only locally. This situation is somewhat similar to that in rapid turbulent shear flow, where the velocities of particles within the same eddy are well correlated, whereas the velocities of particles in different, distant eddies are uncorrelated except in respect to the mean shearing movement in the fluid; it is the local correlation within eddies that is evident in the structure or fabric of the flow. (*Cf.* Townsend, 1956.) It is possible to make a distinction between Sander's direct and indirect componental movement in terms of the range of correlation of the displacement deviations $\delta\mathbf{U}$. If these are uncorrelated for pairs of points at all spacings larger than those comparable with ionic or molecular dimensions, we are dealing with indirect componental movements. However, if the $\delta\mathbf{U}$ are well correlated within regions of dimensions comparable to the grain size or larger, we are dealing with direct componental movements, as in translation gliding or local folding.

Symmetry of kinematic quantities. We have seen that there are several kinematical quantities relevant to the development of fabric; these include the displacements $\mathbf{u}$, the relative displacements $\delta\mathbf{u}$, Gradient $\mathbf{u}$, the pure strain part of Gradient $\mathbf{U}$, and Gradient $\delta\mathbf{U}$. The spatial field of any one of these quantities defines an aspect of the deformation of which the symmetry can be determined. For example, in the problem of strictly homogeneous simple shear illustrated in Figure 14, the displacement field $\mathbf{u}$ has monoclinic symmetry, whereas the associated pure-strain field has orthorhombic symmetry (the displacement field is assumed to be constant in the z direction). However, in this example, $\delta\mathbf{U}=0$ everywhere, and no fabric features can be expected to develop.

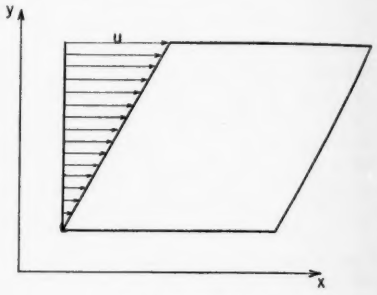

Figure 14. Displacements in strictly homogeneous simple shear

Since the study of fabric can be expected to give direct information only about Gradient $\delta\mathbf{U}$ and the pure-strain part of Gradient $\mathbf{U}$ (which may be written (Gradient $\mathbf{U}$)$_{\text{strain}}$), we need consider only the symmetry of these quantities separately and in combination. Since we are dealing with statistically homogeneous situations, it is sufficient to consider only the point symmetry (the space group is obtained by adding arbitrary translations in all directions). The point symmetry is the same as the symmetry of any spherical projection of a representative sample of the field.

In a representative sample, Gradient $\delta\mathbf{U}$ varies from point to point in a way that is most conveniently represented by the orientation and shape of strain ellipsoids associated with each point (Fig. 13); the orientations of the

local strain perturbation ellipsoids, relative to the mean strain ellipsoid, represent the local rotations. The symmetry of $\delta\mathbf{U}$ can therefore be derived from a spherical projection of the orientations of these strain ellipsoids in the same way as fabric symmetry is determined, but taking into account the shape of the ellipsoids as well. The following symmetries are possible:

1. *Spherical* $K_{\infty h}$ —random distribution of local rotations about all possible axes
2. *Axial* $D_{\infty h}$ —distribution of local rotations relative to the unique axis is independent of azimuth
3. *Orthorhombic* D_{2h}—local rotations distributed symmetrically about the principal axes of the mean strain ellipsoid
4. *Monoclinic* C_{2h} —local rotations distributed symmetrically about one principal axis and asymmetrically about the other principal axes of the mean strain ellipsoid
5. *Triclinic* C_i —local rotations distributed asymmetrically about all principal axes of the mean strain ellipsoid

The symmetry of $(\text{Gradient } \mathbf{U})_{\text{strain}}$ is simply the symmetry of the mean strain ellipsoid, which may be spherical, axial, or orthorhombic. The symmetry elements of Gradient $\delta\mathbf{U}$ will coincide with those of $(\text{Gradient } \mathbf{U})_{\text{strain}}$ in the spherical, axial, and orthorhombic cases. Therefore the possible symmetry groups for the sum of $(\text{Gradient } \mathbf{U})_{\text{strain}}$ and Gradient $\delta\mathbf{U}$ are just those already listed for Gradient $\delta\mathbf{U}$. These are therefore the possible symmetry groups for that part of the "movement plan" that is directly relevant to the final fabric.

An alternative way of representing the deformation, from which the symmetry of the aspects relevant to the fabric can be seen, is by the actual figure into which an initial sphere is distorted (Fig. 13*b*). In a statistically homogeneous transformation, this is approximately an ellipsoid (the strain ellipsoid representing the pure strain part of Gradient $\mathbf{U}$), but, in detail, the figure departs from a true ellipsoid in a way that reveals the local inhomogeneities in the transformation, as can be seen in Figure 13*b*. The symmetry of the actual distorted sphere, viewed on the same scale as the fabric elements, is the symmetry of the part of the "movement picture" that is directly relevant to the final fabric. Thus the symmetry of the distorted sphere in Figure 13*b* is seen to be monoclinic.

A given statistically homogeneous deformation may be composed on a small scale of domains of locally homogeneous deformation in many different ways. For example, the same strain as that in the model of Figure 13 could have been achieved by conjugate shearing in two sets of discrete parallel shear surfaces ("zweischarige Gleitung" as distinct from "einscharige Gleitung"). In this case an analysis of the local perturbations will show that they are distributed symmetrically about all principal axes of the mean strain ellipsoid, whereas in Figure 13 the local perturbations have only one plane of symmetry (the plane of the diagram). In the former case the deformation is orthorhombic, in the latter monoclinic. It is important to note that this symmetry is independent of the presence or absence of the rigid-body rotation of the mean strain ellipsoid. Thus there is no direct relationship between strictly homogeneous "simple shear" and "monoclinic movement" or, conversely, between "pure shear" and "orthorhombic movement," as many geolgists have suggested.

Where only indirect componental movements are involved, the definition of Gradient $\delta\mathbf{U}$ is of no significance. If the $\delta\mathbf{U}$ are completely uncorrelated, this aspect of the deformation has spherical symmetry, and the possible symmetry groups for that part of the deformation that is directly relevant to the final fabric are those for $(\text{Gradient } \mathbf{U})_{\text{strain}}$ —namely, spherical, axial, and orthorhombic.

Dynamic quantities. A force can be represented by a vector acting in a given line. Thus a homogeneous field of force, in which the same force acts at every point, is a homogeneous vector field of point symmetry $C_{\infty v}$. A uniform gravitational field has this symmetry. However, in general, bodies of rock are to be regarded as being subject not to uniform force fields but rather to systems of forces in equilibrium; otherwise they would necessarily be accelerating. Such systems of forces do not constitute homogeneous fields and have only point symmetry.

Within the body of rock, the dynamical quantity that is of direct interest is the stress field, to which the application of the system of forces has given rise. However, stress is not a vector but a tensor quantity of the same nature

as strain. A homogeneous stress field can therefore have spherical ($K_{\infty h}$), axial ($D_{\infty h}$), or orthorhombic (D_{2h}) point symmetry.

In the complete description of the deformation of the rock, the stress–strain relations must also be stated. These will depend in detail on the mechanism of deformation, and their formulation will involve the physical properties of the rock. Therefore, the symmetry of these physical properties must be taken into account in symmetry arguments involving stress–strain relations. However, the possible symmetry groups will depend on the particular mechanism of deformation—that is, on the particular physical properties involved—and will not be discussed further here. For example, if the stress–strain relation were that for perfect elasticity, the elastic properties are concerned, and the possible symmetry groups for these are well known (Voight, 1928).

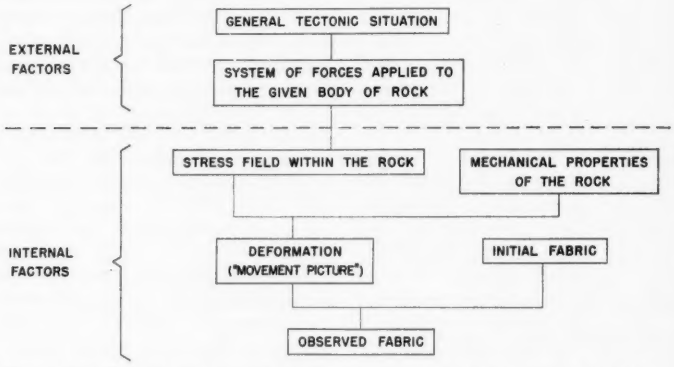

Figure 15. External and internal factors concerned in deformation of a body

Interpretation of Tectonite Fabrics

General. Figure 15 shows how the various factors concerned in the deformation of a body of rock are related (these factors have already been mentioned in the introduction to this Part).

Sander (1930, p. 29–31) has distinguished between the situation in which a deformation obliterates all geometrical features of an initial fabric—a process termed *Umprägung*—and that in which the deformation merely overprints the geometrical features of an initial fabric, leaving them still discernible—a process termed *Überpragung*. In the former case, the structural features of the final fabric can be directly correlated with the deformation and are likely to be relatively simple geometrically; however, the absence of transformed structures hinders the complete kinematical interpretation. On the other hand, fabrics developed by overprinting contain structural features inherited from a previous fabric, as well as new features that have arisen during the deformation, leading in many cases to great complexity. Fabrics developed by overpointing are therefore more fruitful subjects for structural analysis since they carry more of the record of the rock's history. They also are the more common.

Thus the final fabric generally contains fabric elements of the following kinds:

(1) *Inherited fabric elements*, elements inherited in a transformed but recognizable condition from the initial fabric

(2) *Imposed fabric elements*, new features that have arisen entirely as a result of the deformation

(3) *Composite fabric elements*, those that arise from the combination of inherited and imposed features.

A common simple example in which all three kinds of fabric elements may be recognized is illustrated in Figure 16. Sedimentary bedding (inherited fabric element) is folded and crossed by an axial-plane foliation (imposed fabric element); the intersection of these two sets of surfaces is marked by a lineation (composite fabric element). However, it is not always easy to assign given fabric elements to a particular category, especially in the case of preferred orientation of the crystallographic axes of

mineral grains. In the experimental deformation of Yule marble, Turner, Griggs, and coworkers (Turner, 1957, p. 12–16) have found that an initial pattern of preferred orientation of c axes of calcite is progressively modified during deformation, but the final pattern always shows features traceable to the initial pattern, even after large strains.

and Brace (1960) for preferred orientations of crystallographic fabric elements.

We have indicated that the system of forces applied to a given body of rocks can be derived only when the stress field in the rock is understood; that is, forces are more indirectly related to the fabric than are stresses. However, in an inhomogeneous body, where we are concerned with a particular structural feature in isolation (that is, in a nonpenetrative form) rather than as an element in a statistically homogeneous fabric, for example, a single fold, it may be profitable to attempt direct correlation with the system of forces applied to the region. Any symmetry arguments in such an analysis will involve only point symmetries of the features and the system of forces, rather than spatial symmetries as in the case of fabric.

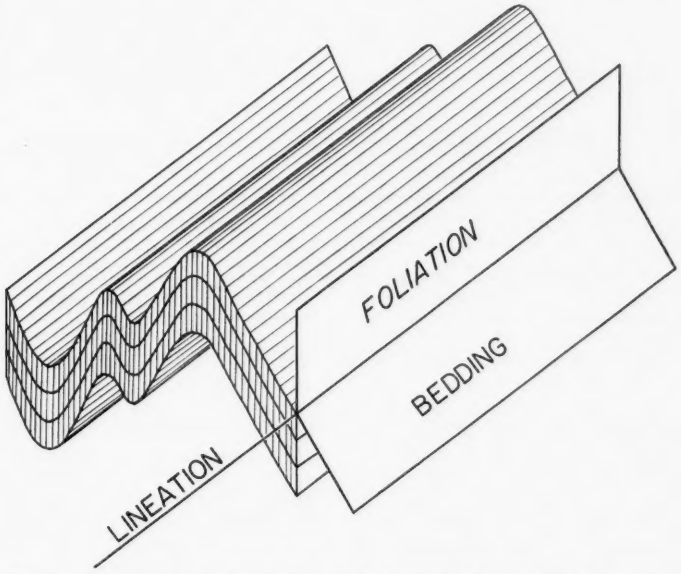

Figure 16. Examples of inherited (bedding), imposed (foliation), and composite (lineation) fabric elements

In further interpretation of the fabric, particular fabric features are related to particular factors in the rock's history. As indicated, the most direct correlation is usually with some aspect of the deformation and with the initial fabric. Such a purely geometrical interpretation is termed kinematic interpretation. It is our main concern in this paper and is dealt with in more detail in the next section. However, in some cases where indirect componental movements predominate (especially if the rock has deformed by creep processes involving mainly diffusion), the "movement picture" itself is of little interest. In these cases, it is simpler to attempt to correlate the fabric directly with the stress field, as is done in the theories of Kamb (1959), MacDonald (1960),

Fabric as the imprint of the deformation. We have shown that in any statistically homogeneous region the deformation, referred to internal co-ordinates, can be resolved into two parts: (1) a pure strain (the mean strain), which is everywhere the same, and (2) a local strain accompanied by a rotation that can vary

from point to point. Particular aspects of the fabric may bear the imprint of one or other of these components of the deformation, and it is an important part of interpretation to classify the observed fabric elements accordingly.

(1) *Noncrystallographic fabric elements.* If fabric features have been mechanically ineffective—that is, if they do not correspond to or give rise to local inhomogeneities in the deformation—we shall call them *kinematically passive.* Kinematically passive noncrystallographic fabric elements are always inherited fabric elements in the sense defined earlier. They are lines or surfaces of material points (including boundaries of local regions such as pebbles), which may be recognized before and after the deformation by virtue of visible characteristics such as color or minor peculiarities in mineralogy. They therefore act as markers within the body and are transformed geometrically ("internally rotated") from their initial configuration in accordance with the deformation. In principle, they contain a complete record of the deformation; their mean configuration reflects the mean strain, and their minor distortions reflect the local inhomogeneities of the deformation (for example, slip folding of bedding or foliation).

Not all inherited noncrystallographic fabric elements are kinematically passive. If they are the sites of inhomogeneities of deformation, we shall call them *kinematically active.* They are then to be interpreted as directly reflecting the inhomogeneities in deformation. However, the kinematically active features may not be tied to the same material points throughout the deformation and therefore need not act as markers in reflecting the mean strain. We therefore distinguish two cases for kinematically active inherited elements: if they act as markers, they reflect both components of the deformation; if they do not, they directly reflect only the local inhomogeneities of deformation.

Imposed noncrystallographic fabric elements are always initially sites of inhomogeneity of deformation and therefore are always kinematically active. They may become kinematically passive where later generations of imposed elements appear. In general they do not act as markers and so can be regarded only as bearing a record of the local inhomogeneities of deformation. Composite fabric elements will have the same properties as the imposed features from which they have partly arisen.

In Figure 17 these relationships are summarized for noncrystallographic fabric elements.

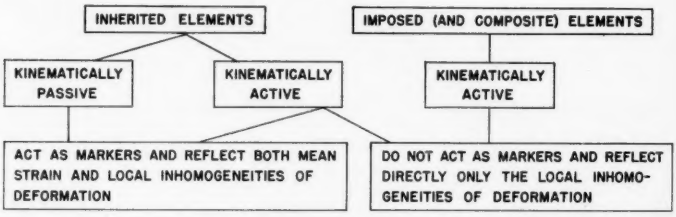
INHERITED ELEMENTS
IMPOSED (AND COMPOSITE) ELEMENTS
KINEMATICALLY PASSIVE
KINEMATICALLY ACTIVE
KINEMATICALLY ACTIVE
ACT AS MARKERS AND REFLECT BOTH MEAN STRAIN AND LOCAL INHOMOGENEITIES OF DEFORMATION
DO NOT ACT AS MARKERS AND REFLECT DIRECTLY ONLY THE LOCAL INHOMOGENEITIES OF DEFORMATION

Figure 17. Functions of noncrystallographic fabric elements during deformation

(2) *Crystallographic fabric elements.* A deformation that is statistically homogeneous on scales coarser than the grain size can influence the preferred orientation of the crystallographic fabric elements in two ways: (1) by introducing a preferred orientation where none existed previously; (2) by modifying an existing preferred orientation. In either case, the preferred orientation will bear the imprint of both components of the deformation. Any attempt to separate the roles of the two components will involve elucidating the mechanism of deformation on the scale of the grains (including such questions as the extent to which the strain in individual grains is the same as the mean strain in the aggregate, whether the grains themselves are unstrained, and if so whether they are rotating in the deforming matrix,[16] or whether particular gliding mechanisms are active within the grains). We shall not investigate these questions here, so we can apply symmetry arguments only to the relation of the preferred orientation pattern to the

[16] The study of rolled garnets may throw light on this question.

over-all deformation. The features revealed by A.V.A. studies (Sander, 1950, p. 39) can be classified as noncrystallographic fabric elements.

Symmetry principles in interpretation. In the interpretation of the observations on the fabric, it is in general impossible to deduce quantitatively the full history of the body of rock, although in favorable cases some progress may be made (as in the study of deformed pebbles or fossils for example). Therefore, we fall back on symmetry arguments, as explained in the Introduction, and in this final section we shall attempt to state the general principles underlying interpretations in terms of symmetry.

The application of symmetry arguments in the interpretation of deformed rocks was pioneered by Sander and Schmidt and was based on the general hypothesis that the symmetry of the physical factors involved in the deformation is reflected in the geometrical fabric (*cf.* Sander, 1948, p. 3: ". . . dass sich die Symmetrie vektorieller Gefüge . . . im morphologischen Gefüge abbildet"). In kinematic analysis, Sander's principle is usually expressed in a form such as the following: "The symmetry of a tectonite fabric reflects the symmetry of the movement plan of deformation" (Turner and Verhoogen, 1960, p. 628). We have already commented on the resemblance between Sander's principle and some of Curie's statements about symmetry in physical phenomena in general. In the following we shall restate Curie's principles in terms directly applicable to structural analysis and show how these embrace Sander's principle and clarify certain aspects of it. *Statistical homogeneity is assumed in all cases.*

In the kinematic interpretation of a tectonite fabric, the initial fabric and the two components (mean strain and local perturbations) of the deformation referred to internal axes are the independent contributing factors or "causes," and the final fabric is the "effect." By the principle of superposition, the resultant symmetry of the contributing factors is the symmetry that is common to the initial fabric and to the two components of the deformation. Applying Curie's principles further, *this resultant symmetry must then appear in the final fabric.* (*Cf.* Sander's principle.) *Any symmetry elements absent in the final must have been absent in the initial fabric or in one of the deformation components. However, the final fabric may possess symmetry elements that are absent in the initial fabric or the deformation.*

For example, if the initial fabric were spherically symmetrical (symmetry $K_{\infty h}$) and the over-all symmetry of the deformation were monoclinic, the resultant symmetry of the "causes" would be monoclinic—that is, they would possess a reflection plane and a twofold rotation axis normal to it, repeated by arbitrary translations in all directions. The final fabric must then have a point symmetry consisting of the same reflection plane and twofold rotation axis. This requirement would be met by a fabric consisting of a single foliation surface in the appropriate orientation. However, such a fabric, which is of axial point symmetry ($D_{\infty h}$), also has reflection planes and an infinitefold axis normal to the foliation plane, which is allowed by Curie's rules. On the other hand, the absence of a reflection plane at 45° to the foliation, for example, can be taken as evidence of the absence of such a reflection plane in either the initial fabric or the deformation.

These symmetry arguments may be applied to individual subfabrics. In this case, the subfabric of a particular feature will have a symmetry that will include at least the symmetry elements common to the symmetry of this feature in the original rock (if the feature existed initially) and to that of the particular components of the deformation that have influenced it; any symmetry elements absent in the final subfabric must have been absent in either the initial fabric or the relevant components of the deformation. By applying the principle of superposition, the same symmetry arguments may be extended to the complete fabric and all the aspects of the initial fabric and deformation related to it.

If a particular subfabric is absent or has complete spherical symmetry in the initial fabric, its symmetry in the final fabric will contain all the symmetry elements of the aspect of the deformation that influenced it. We therefore see that imposed fabric elements are of special importance in the interpretation of the deformational history of a rock since the symmetry information they bear can be directly related to the deformation. Moreover, we have seen that only the local perturbation component of the deformation is directly related to the imposed fabric elements. Therefore, according to Curie's principles, the *local perturbation component of the deformation cannot have more symmetry than that revealed in the imposed fabric elements, although the latter may contain symmetry elements absent in the deformation component; however, any symmetry of this deforma-*

tion component must appear in imposed subfabrics to which it gives rise.

On the other hand, the symmetry of subfabrics of inherited features cannot be correlated directly with the symmetry of deformation since they are also influenced by the symmetry of initial fabric. As a result, inherited subfabrics (especially those of kinematically passive features) are frequently of lower symmetry than imposed subfabrics. (*See*, for instance, Weiss and McIntyre, 1957, p. 599, 600.) Also, owing to the influence of the initial fabric, the symmetry of fabrics containing inherited elements (especially kinematically passive) will in general be lower than the symmetry of the deformation.

So far we have discussed the symmetry arguments that can be applied when it is possible to interpret particular fabric features as being related to particular aspects of a deformation. When such relationships cannot be established, symmetry arguments enable one to conclude that *whatever the nature of the contributing factors, the symmetry that is common to them cannot be higher than the symmetry of the observed fabric, and symmetry elements absent in the fabric must be absent in at least one of the contributing factors.* In kinematic interpretation, such an argument provides information about the initial fabric and the deformation. In other cases as, for example, when a preferred orientation of crystallographic fabric elements is to be attributed directly to the influence of a stress field, the same symmetry arguments may be applied to deduce the symmetry common to the stress field and any initial preferred orientation.

Similar applications of these symmetry principles may be fruitful in other branches of geology also, as for example in interpreting the fabrics of sedimentary rocks, both those arising by direct deposition and those arising by diagenetic processes. Sander (1948; 1950) has discussed the role of symmetry in a number of geological problems other than those of deformed rocks. When physical properties such as magnetic, elastic, or electrical properties are measured, the same symmetry principles can be applied here too to gain some information about the circumstances in which these properties were acquired.

The symmetry of a bulk physical property of a rock is likewise a guide to symmetry of its fabric (functional fabric). Neumann's principle states that the symmetry element of any physical property of a crystal must include all the symmetry elements of its geometrical (morphological) symmetry. Where this principle is applied to aggregates, the study of easily determinable physical properties (such as bulk electrical, magnetic, or thermal properties) may be a convenient guide to the geometrical symmetry of its fabric by showing which symmetry elements are lacking. Further, the measurement of properties such as piezoelectricity allows detection of the absence of a center of symmetry, not otherwise possible. Such investigations are likely to be most useful in determining fabric symmetry in very weakly deformed rocks where conventional methods of geometrical analysis yield conclusions of doubtful statistical significance, and in very fine-grained rocks in which conventional microscopic study is impossible.

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Exploration Geophysics: A Review

Abstract: The art of exploration geophysics is reviewed with reference to classical and current literature. Its main features are described, the state of instrumentation and theory in it is discussed, and the problems it has yet to solve are examined.

Exploration geophysics is both a science and an industry. As a science it draws on its parent sciences of physics and geology, and it has a reciprocal relationship with its sister disciplines of seismology and geodesy. Since its predictions often cannot be verified, it lacks one indispensable ingredient of science —the power of self criticism. It also suffers because its discoveries are proprietary and therefore often not published to be built on or challenged. In spite of these handicaps, its scientific status is rising. More trained workers than ever before are in the field, and their publications are of better quality.

As an industry, exploration geophysics has long had the financial support that public funding has only recently given to science in general. It has served the demand for mineral discoveries well, but has sometimes failed to give the best possible support to its clients through lack of rapport between those who study problems and those who make decisions about them.

The instruments and devices used by exploration geophysics have in general surpassed its fundamental needs. In 1930 the reflection seismograph was untested, and the field gravity meter was not invented, although they were clearly needed, and their prototypes, the refraction seismograph and the torsion balance, were plainly inadequate. Now it is possible to take almost any geophysical measurements found desirable, and the instrumental problem is usually to take them faster and with lighter and more dependable gear.

The present frontiers of the art are not in the observing of data but in their interpretation and application to the search for minerals. The interpretation of seismic records is still in a relatively primitive form. How far it will be developed in the future depends on how acutely it is needed, but it is not easy to see the end of the data-processing techniques that are now being studied. The interpretation of gravimetric, magnetic, electrical, and radioactivity observations still requires personal skill, although automatic computing techniques have been applied to the first two types.

Résumé: L'auteur passe en revue l'art de l'exploration géophysique d'après les travaux classiques et contemporains. Il en décrit les principaux caractères, discute de l'état actuel de l'instrumentation et de la théorie et examine les problèmes qui restent à résoudre.

La géophysique d'exploration est à la fois une science et une industrie. Comme science elle s'appuie sur les sciences mères de la physique et de la géologie, et à des rapports étroits avec les disciplines soeurs de la séismologie et de la géodesie. Parceque ses prédictions ne peuvent pas souvent être vérifiées, il lui manque un élément indispensable à la science, le pouvoir de se critiquer soi-même. Elle souffre aussi du fait que ses découvertes sont soumises aux droits de la propriété, et ne sont donc pas généralement rendues publiques, donc ne peuvent pas être utilisées, ou mises en question. Malgré de tels handicaps, son rang dans le monde scientifique s'élève. De plus en plus de chercheurs exercés s'y adonnent et la qualité de leurs publications s'améliore.

En qualité d'industrie, la géophysique d' exploration reçoit depuis longtemps l'appui financier que les finances publiques n'ont que récemment accordé à la science en général. Elle a rendu de grands services dans la découverte de ressources minérales, mais n'a pas toujours pu donner le meilleur appui à cause des divergences de vue qui séparent ceux qui étudient les problèmes de ceux qui prennent les décisions à leur sujet.

Les instruments et techniques utilisés par la géophysique d'exploration ont en général devancé ses besoins fondamentaux. En 1930, le séismographe à réflexion n'avait même pas été mis à l'épreuve et les appareils de mesure du champ de la pesanteur n'étaient pas inventés, et pourtant le besoin s'en faisait sentir, et leurs prototypes, le séismographe à réfraction et la balance à torsion, étaient manifestement insuffisants. De nos jours on peut faire presque n'importe quelle mesure géophysique qui soit souhaitable, et le problème des instruments se réduit à accélérer les mesures et à alléger et rendre plus précis les appareils.

Les limites actuelles de l'art ne résident pas dans l'observation des faits, mais dans leur interprétation et dans leur application à la recherche des minéraux. L'interprétation des données séismiques est encore à un stade assez primitif. Son développement futur dépendra de l'urgence des besoins, mais on voit mal où aboutiront les techniques d'analyse des données actuellement à l'étude. L'interprétation des données

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d'observation gravimétriques, magnétiques, électriques et radioactives demande encore un talent personnel, quoique des techniques mécanographiques de calcul commencent à être employées dans les 2 premiers cas.

Resumen: Se hace una reseña del arte de la geofísica exploratoria con referencia a la literatura clásica y actuel. Se describen sus aspectos principales, se discute el estado de la instrumentación y de la teoría y se examinan los problemas que se tienen aún por resolver.

La geofísica exploratoria es tanto una ciencia como una industria. Como ciencia, depende de las ciencias originadoras, como la física y la geología, y tiene relaciones recíprocas con las disciplinas hermanas, que son la sismología y la geodesia. Debido a que sus pronósticos a menudo no pueden ser comprobados, carece de un ingrediente indispensable de una ciencia, o sea la fuerza de la autocrítica. Sufre también la inconveniencia de que sus descubrimientos son de naturaleza privada y pocas veces se publican, evitando así que se aprovechen en trabajos nuevos o se critiquen. A pesar de estas desventajas, su posición científica está subiendo. Un número mayor que nunca de investigadores capacitados se encuentran en esta profesión, y sus publicaciones son de mejor calidad.

Como una industria, la geofísica exploratoria ha tenido por largo tiempo el apoyo económico que sólo recientemente han dado las instituciones gubernamentales a las ciencias en general. Ha servido bien a la demanda de descubrir yacimientos minerales, pero a veces no ha dado el mejor apoyo posible a sus clientes, por falta de una colaboración estrecha entre los que estudian los problemas y los que toman las decisiones sobre ellos.

Los instrumentos y equipo usados en la geofísica exploratoria en general, han sobrepasado las necesidades fundamentales. En 1930, el sismógrafo de reflección no se había probado y el gravímetro de campo no se había inventado, aunque eran altamente necesitados, y sus prototipos, el sismógrafo de refracción y la balanza de torsión, eran claramente inadecuados. Actualmente es posible tomar casi cualquier medición geofísica que se desee, y el problema instrumental generalmente consiste en reducir el tiempo necesario para tomar la mediciones, disminuir el peso del equipo y hacerlo más seguro.

Las fronteras actuales de este arte no se encuentran en la observación de los datos, sino en su interpretación y aplicación en la búsqueda de minerales. La interpretación de registros sísmicos está aún en una forma primitiva. Qué tan lejos se desarrollará esta técnica en el futuro dependerá de la urgencia con que se necesitará, pero no es fácil prever el fin del desarrollo de las técnicas empleadas para procesar los datos que actualmente se encuentran en estudio. La interpretación de observaciones gravimétricas, magnéticas, eléctricas y de radioactividad aún requieren la pericia personal, aunque ya se han empleado para los dos primeros tipos de observaciones, técnicas de cómputo automático.

Zusammenfassung: Die Anwendbarkeit der Erforschungs-Geophysik wird mit Bezugnahme auf die klassische und gegenwärtige Literatur besprochen. Ihre Hauptmerkmale werden beschrieben, der Entwicklungsstand der Instrumente und ihre theoretischen Grundlagen werden diskutiert, und die Probleme, die noch zu lösen sind, untersucht.

Erforschungs-Geophysik ist beides Wissenschaft und Industriezweig. Als Wissenschaft fusst sie auf ihren elterlichen Wissenschaften, Physik und Geologie, und steht in enger Beziehung zu ihren Schwester-Fächern, der Erdbeben-und der Vermessungskunde. Da ihre Voraussagungen nicht oft bestätigt werden können, fehlt ihr ein für die Wissenschaft unerlässlicher Bestandteil, der Einfluss der Selbstkritik. Sie leidet auch darunter, dass ihre Entdeckungen dem Eigentümerrecht unterfallen und nicht oft veröffentlicht werden, sodass man darauf aufbauen oder dagegen Einwendungen machen könnte. Trotz dieser Hindernisse ist ihre wissenschaftliche Anerkennung im Aufstieg begriffen. Mehr ausgebildete Arbeitskräfte als je zuvor sind auf dem Gebiet tätig, und ihre Veröffentlichungen sind von besserer Qualität.

Als Industriezweig gesehen hat die Erforschungs-Geophysik seit langem die finanzielle Unterstützung gehabt, die erst seit kurzem aus öffentlichen Geldmitteln für die allgemeine Wissenschaft zur Verfügung gestellt wurde. Sie hat, was die Entdeckung von Mineralien anbelangt, der Anforderung voll genügt, aber sie hat manchmal versäumt, ihren Klienten die bestmögliche Unterstützung zu geben. Das ist eine Folge von mangelnder Verbindung zwischen denen, die Probleme untersuchen und denen, die Entscheidungen darüber treffen. Die Instrumente und Vorrichtungen, die von der Erforschungs-Geophysik benutzt werden, haben im allgemeinen ihre fundamentalen Bedürfnisse übertroffen. Im Jahre 1930 war der Reflektions-Seismograph noch nicht ausprobiert, und der statische Schweremesser noch nicht erfunden, obwohl beide eindeutig gebraucht wurden, und ihre Vorbilder, der Refraktions-Seismograph und die Torsionswaage, deutlich unzureichend waren. Heute ist es möglich, fast alle geophysikalischen Messungen vorzunehmen, die wünschenswert erscheinen, und das Problem ist gewöhnlich, solche Instrumente zu verwenden, die Messungen schneller und mit leichterem und zuverlässigerem Zubehör vornehmen.

Die gegenwärtigen Grenzen der Erforschungs-Geophysik liegen nicht in der Beobachtung von Werten, sondern in deren Auswertung und Anwendbarkeit für die Suche nach Mineralien. Die

Auswertung von seismischen Registrierungen geschieht noch immer in einer verhältnismässig primitiven Art. Wie weit sie in Zukunft entwickelt werden wird, hängt davon ab, wie dringend sie gebraucht wird. Es ist jedoch nicht einfach, das Ende der auswertenden Methoden, an denen jetzt gearbeitet wird, abzusehen. Die Auswertung von gravimetrischen, magnetischen, elektrischen und radioaktiven Beobachtungen erfordert noch immer persönliche Geschicklichkeit, obwohl automatische Berechnungsmethoden auf den beiden erst genannten Gebieten angewandt worden sind.

Ф. Е РОМБЕРГ

Разведочная Геофизика (обзор)

Резюме: Автор рассматривает разведочную геофизику на основании классической и текущей литературы. Дана общая характеристика геофизики, состояние оснащения инструментами, теоретические основы, и рассматриваются вопросы, которые еще предстоит разрешить при помощи геофизических методов разведки.

Разведочная геофизика объединяет в себе науку и прикладное искусство. В качестве науки, геофизика базируется на физике и геологии, и обычно связана с родственной сейсмологией и с геодезией. Ввиду того, что зачастую нельзя проверить выводов сделанных по данным геофизических методов, геофизика лишена возможности критического анализа, играющего неотъемлимую роль в науке. Недостатком геофизики является также то, что геофизические открытия делаются собственностью предприятий, не всегда опубликовываются и не могут служить основой для дальнейших исследований, равно, как и не могут никем оспариваться. Несмотря на эти помехи, значение геофизики, как науки, растет. В этой области сейчас работает больше чем когда либо хорошо подготовленных специалистов и качество их печатных работ соответственно повышается.

В производственном отношении разведочная геофизика, с давних времен, имела финансовую поддержку, в той степени, в которой, за счет общественных средств, только недавно стала оказываться помощь науке. Геофизики хорошо справились с задачей открытия новых месторождений полезных ископаемых, но иногда они не умели, в достаточной степени, оказать поддержку заказчикам, вследствие недостатка взаимопонимания, которое может возникать между исполнителями и администрацией.

Инструменты и оборудование, которыми пользуются в разведочной геофизике, в общем, переросли основные требования предъявляемые к ним. В 1930 г. сейсмограф для работ с отраженными волнами еще не подвергался испытанию, а полевой гравиметр еще даже не был изобретен, хотя в этих инструментах ощущалась острая нужда. Теперь же имеется возможность производить любые геофизические измерения при помощи более легкого и надежного оборудования. Поле деятельности геофизиков заключается теперь не только в наблюдениях но и в толковании данных съемки и их применении при поисках полезных ископаемых. Развитие интерпретации записей сейсмической разведки еще находится на относительно примитивном уровне. Дальнейшее развитие этого метода зависит от того, в какой степени он будет находить применение. Сейчас трудно предвидеть конечную стадию развития методов интерпретации геофизических данных. Эти методы сейчас изучаются. Интерпретация данных гравиметрической, магнитной, электроразведочной, и данных радиометрической съемки, до сего времени требуют личного участия и квалификации геофизика, хотя автоматические счетные машины уже применяются для интерпретации гравиметрических и магнитных методов геофизической разведки.

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INTRODUCTION

Methods

Exploration geophysics is the art of applying the techniques of geology and physics to the discovery of petroleum and certain minerals. Economic geology asks the questions, and physics answers them. The task of the exploration geophysicist is to look for correlations between the occurrence of minerals and observable physical phenomena.

The most important exploration method is

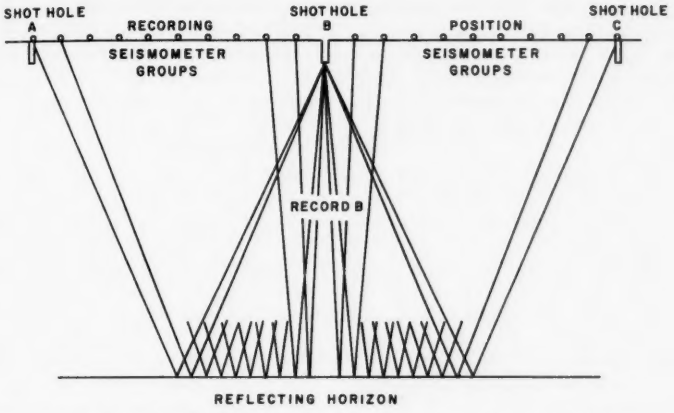

Figure 1. Reflection seismograph; array, reflector, ray paths (Geophysical Service, Inc.)

the reflection seismograph. An array of seismometers is spread out on the ground, and each seismometer connected through an amplifier to the recording element in an oscillograph (Fig. 1). When a charge is exploded in a nearby hole, portions of the energy in the resulting seismic wave are reflected from interfaces and return to the surface. The reflected waves cause the seismometers to move minutely and send signals to the oscillograph, which makes a record showing their arrival (Fig. 2). The depths of the reflecting interfaces are computed from their travel times. From a set of such records cross sections of contour maps showing geologic structure are made.

The refraction seismograph uses a charge at a distance from the array instead of close to it. The seismic wave travels downward to a high-velocity layer, overtakes the energy which travels along the surface, and rises and reaches

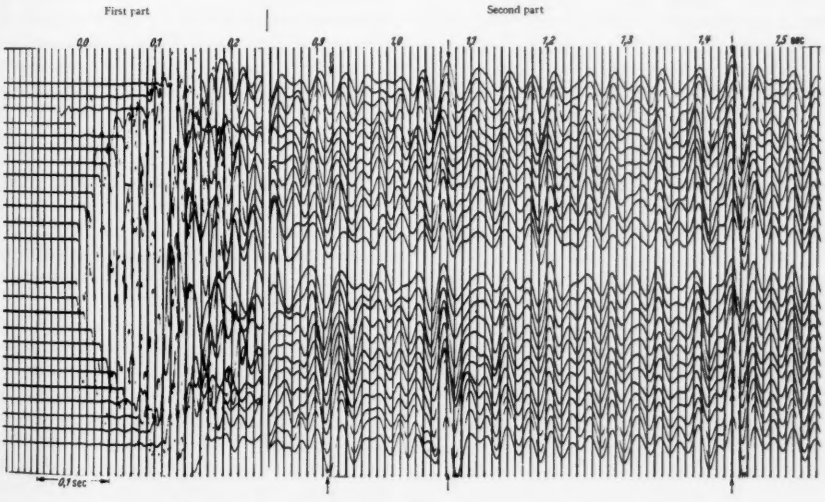

Figure 2. Oscillogram or record of a reflection shot (Geophysical Service, Inc.)

the array (Fig. 3). The depth of the high-velocity layer can then be computed from the arrival time of the wave, and a contour map made from a set of observations as in the reflection method. Refractions are less accurate and are usually used when reflections cannot be observed.

The gravity method consists of measuring the gravitational field of the earth at various points on its surface. After corrections have been made for topography, elevation, and change in latitude, a contour map is made showing the local fluctuations in gravity at the surface. These can be interpreted as being due to geological features such as intrusions of light material like salt or heavy material such as basalt, or to faults or anticlinal or synclinal structures (Fig. 4).

The magnetic method is much like the gravity method. The magnetic intensity of the earth's field is measured at a set of locations, and the values plotted and contoured in the same way as with gravity. Anomalies in the magnetic field are evidence of anomalies in the geologic structure just as gravity anomalies are, except that sedimentary structures give very weak magnetic anomalies, as a rule, and igneous structures give very strong ones (Fig. 5). Magnetometers have been adapted to airborne use, and most magnetic exploration is now done from the air.

The electrical method consists of measuring the voltage between two electrodes in the earth. The voltage may be due either to natural currents or to artificial ones set up for the purpose. Differences in electrical response between one pair of points and another are evidence of differences in resistivity of the earth below them, and these differences have some geologic cause (Fig. 6). Electrical methods are used extensively in bore holes. In electromagnetic or induction methods an alternating current is set up in a wire loop. If a conducting body, such as a body of metallic ore, is present, currents will flow in the body and can be detected with search coils (Fig. 7). Radioactivity is used in searching for ores that are radioactive or that are associated with radioactive elements. Radioactive well logging has proved useful; a sensitive element is lowered into a bore hole, and changes in radioactivity at different depths are used to aid subsurface correlation.

History

The history of exploration geophysics is short. Although the use of a compass needle as a primitive magnetometer has been known for a long time, it was only in the nineteenth century that electrical methods became well enough understood to be used for prospecting. Later the torsion balance was invented, and for the first time gravity was measured accurately enough to find local anomalies. Artificial earthquakes came to be used for exploration only after 1920, and the reflection seismograph only after 1930. The reflection seismograph set off the expansion in oil prospecting which caused the industry to assume the aspect it has today, with exploration parties numbered in the hundreds all over the world, and dozens of research laboratories advancing the art. The field gravity meter came into use in the late 1930's, and the flying magnetometer after 1945. Electrical and radioactivity methods have had a steady growth and advancement but have never become as widely used as seismic, gravity, or magnetic methods.

Publications

The most important publication in exploration geophysics is the periodical GEOPHYSICS published by the Society of Exploration Geophysicists. Its twenty-fifth anniversary number (February 1960) included several review and historical articles, including one by the editor on the classics of geophysics. It has risen greatly in quality and in volume through its life, and it is in itself a fairly good reference library on its subject. Next in importance is GEOPHYSICAL PROSPECTING published by the European Association of Exploration Geophysicists and now in its eighth year. Like GEOPHYSICS, it contains articles on both practical and philosophical subjects with rather a larger proportion of articles on mining.

The principal books on exploration methods are by Nettleton (1940), Heiland (1946), Jakosky (1960), and Dobrin (1960); the last two are earlier works recently revised. Two volumes of case histories, edited by Nettleton (1948) and by Lyons (1956), are part of the classical literature of exploration.

Publications stressing scientific rather than commercial geophysics are the JOURNAL OF GEOPHYSICAL RESEARCH, TRANSACTIONS OF THE AMERICAN GEOPHYSICAL UNION, and the BULLETIN OF THE SEISMOLOGICAL SOCIETY OF AMERICA in the United States, and the GEOPHYSICAL JOURNAL of the Royal Astronomical Society in the United Kingdom. The most important of the non-English publications, the

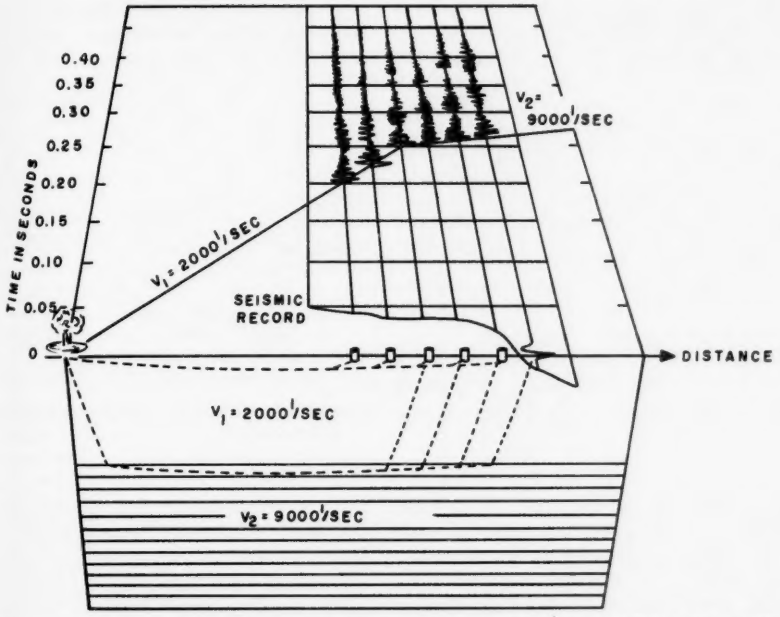

Figure 3. Refraction seismograph: array, refractor, and record (After Heiland, 1946, Figs. 2–10)

Bulletin of the Academy of Sciences, U.S.S.R., Geophysics Series, is fortunately available in translation. A survey of Russian geophysical literature published by Roark (1959) was statistical rather than critical: he showed that about the same bulk of material on the subject was published in Russian as in English. In general the tone of the Russian articles is slightly more academic and fundamental with less reference to applications. They tend to refer freely to articles written in English; the reverse is not true.

In addition to the regular periodicals, numbers of special papers are published by the U.S.

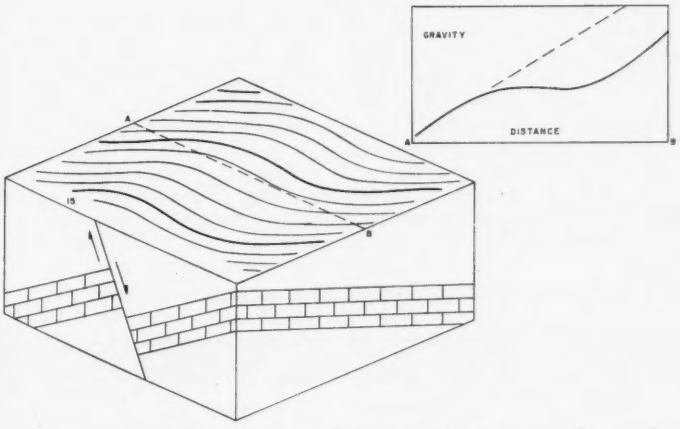

Figure 4. Contours of gravity across a fault

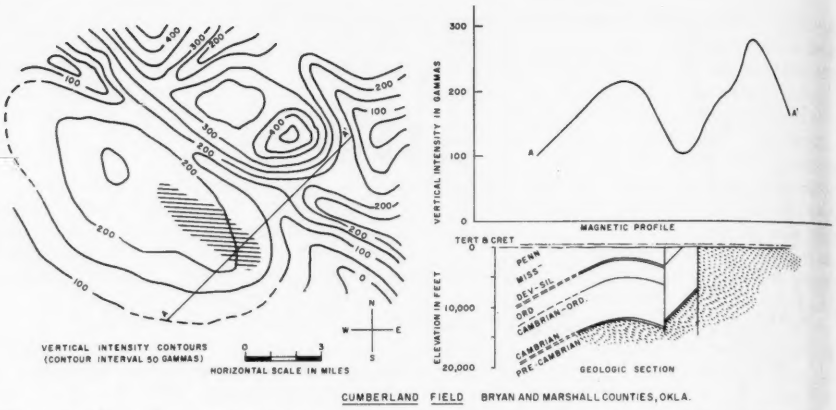

Figure 5. Observed magnetic intensity and structure over Cumberland field, Oklahoma (After J. W. Peters, 1949, Fig. 8)

Geological Survey, the Bureau of Mines, State geological agencies, and universities. The same is true to a lesser extent in other countries. These publications naturally tend to stress scientific rather than commercial applications of geophysical methods but are nevertheless of interest to exploration geophysicists as illustrations and as sources of geological and geophysical information about areas, particular or typical, that may come under observation.

Many reviews and general articles on exploration have appeared recently, for example: Bruckshaw's (1959) presidential address to the European Association of Exploration Geophysicists, a critique of exploration techniques and their probable future. A review of scientific geophysics by Adams (1960) for the International Union of Geodesy and Geophysics contained separate reviews of the branches of geophysics, including one by Freeman Gilbert (1960) on applications of electronic computers to problems in seismology; this is probably an important line of advance for the near future. Enslin's (1955) review of geophysical prospect-

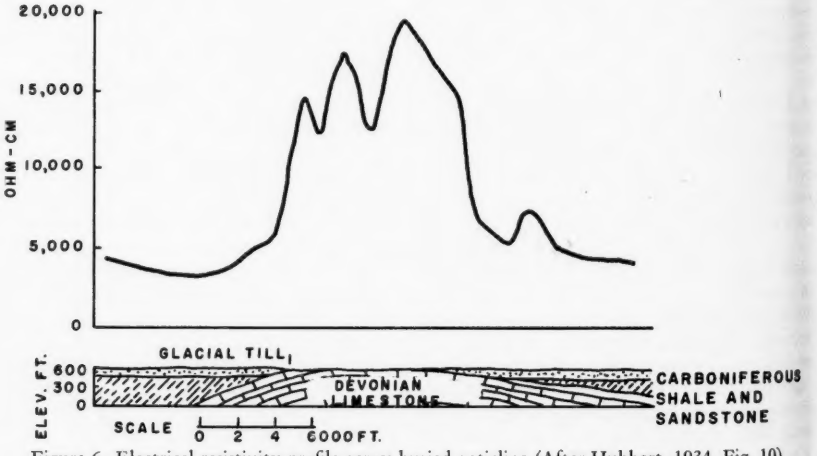

Figure 6. Electrical resistivity profile across buried anticline (After Hubbert, 1934, Fig. 10)

ing in South Africa describes exploration in a subcontinental area in which ground water and gold are more important than oil.

ACKNOWLEDGMENTS

Acknowledgment is due R. C. Dunlap, Jr., and Francis Birch for their criticisms and suggestions. Especial thanks are due Isidore Zietz who noted several important omissions and points where clarification was needed, and above all to Preston Cloud, Jr., who suggested that the review be written and who criticized it in detail.

RELATION TO SCIENCE

General Features of Exploration Geophysics

Exploration geophysics is unique among the new technologies because of its relationship to the experimental science of physics, the descriptive science of geology, and the unpredictable demands of commerce. Geology sets the problems; physics provides the instruments and the interpretive theory, and commerce offers the incentive. The relation among the three is subject to constant change in ways that are foreign to other applied science. Physics and geology are so dissimilar that the relationship between them and exploration is difficult to define. Exploration is therefore different from more straightforward scientific techniques such as the various forms of engineering.

An important hindrance to progress in exploration is its lack of a method for profiting from the body of experience in the field. Some of this is inevitable because commercial organizations tend to keep to themselves the improvements developed at their expense. The chief reason, however, why useful experience is not gained is that exploration campaigns are usually planned to test a particular area for minerals rather than to establish valid general methods. This prevents the explorer from building on a body of knowledge collected and codified by his predecessors, as is done in geology or physics. Another handicap is the tendency to allow and even to encourage exploration to degenerate into prescribed routines. The reason for this tendency is not hard to find. The work of the field party is partly wasted unless it is directed by a senior geophysicist with a sound knowledge of both physics and geology. When field work is expanded there are not enough such people, and the field parties must be guided by routine instructions if they are to operate at all.

The most promising line of advance in exploration is the reduction of data to reproducible form and the automatic processing made possible by such reduction. Automatic processing is now being used to recover signals hidden by high-amplitude noise. Since the level of seismic noise in the earth is the ultimate barrier to the reception of signals, improvement must be based on enhancing signal level with respect to noise level. Graebner (1960) showed how to adapt anti-noise techniques to local differences in noise character. In addition to allowing advanced methods of noise reduction, automatic processing will permit a new departure in interpretation. So far only visual criteria have been used to interpret seismic data. This leads to ambiguity because it is hard to keep the criteria objective. Development of automatic operators to match the signals to the criteria would make the process objective and open the way to improvement. Rieber (1937) invented a device to apply the criterion of alignment to signals on records, but his lead was not followed.

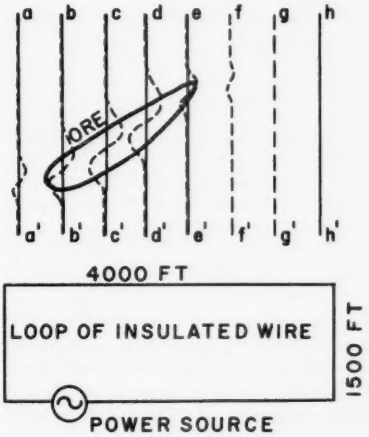

Figure 7. Detection of conducting ore body by electromagnetic induction. Search coils are moved along cross line and give an anomalous response near the ore body (After Dobrin, 1960, Figs. 17–19)

Many suggestions have been made for other avenues of advance in seismology. Press (1957) showed a relation between geologic structure and phase velocity. Knopoff (1959) made calculations showing a relation between structure and scattering. Crawford and others (1960)

described a steady-state source of seismic energy. So far little use has been made of the shape of the seismic signal.

Advances in gravity and magnetic methods are more likely to be made in interpretive skill than in collecting and processing data. Even though faster and steadier instruments may be built, those now available are accurate enough to be limited by the noise level. Electrical and radioactive methods are in active development, including a method of measuring the plane of polarization of magnetic noise. These developments will affect mining more than oil exploration.

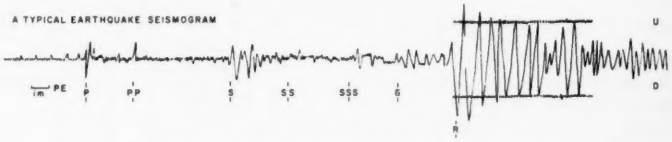

Figure 8. Seismogram of a natural earthquake, showing different types of waves

Physics

The relation of exploration geophysics to physics itself is straightforward. Exploration instruments from the compass needle to the digital computer are all in the domain of physics and were developed, except in their most primitive form, by physicists or electrical engineers.

The interpretation of geophysical data in terms of geologic information uses both physics and geology. Attempts have been made to divide the process so that physicists observe and treat the data and geologists decide what they mean. The division is artificial because the relation between geophysical data and geologic structure is not subject to codifying. Exploration is continually moving to new areas and trying out new methods, so that rules made for solving one problem do not apply to the next. Solving each new problem calls for a continuous interchange of ideas between physics and geology; the interpreter must understand the physical relation between observations and geologic structure, the importance of the probable errors in his data, and the degree of resolution in his methods; above all he must know when the interpretive process moves from logic to speculation.

Seismology

Exploration geophysics and scientific seismology are closely interdependent. Seismographs were originally developed for studying earthquakes and were later adapted for commercial use, but since then earthquake seismographs have benefited from improvements in exploration gear. The study of natural earthquakes is the parent science of exploration seismology. The chief difference between the two is that the theory of earthquakes recognizes and describes the different kinds of seismic waves (Fig. 8), while in exploration only the head waves are considered. This points to a possible direction of progress. Gutenberg (1956) discussed what happens to waves leaving bedrock and traveling through alluvium; this is important because reflection seismometers must be placed on alluvium or weathered rock and are not useful when planted on hard rock.

Seismology has been slow to take advantage of the techniques developed by exploration for convenience in collecting and processing data. However, the recent impetus given to seismology by political developments has accelerated the conversion to automatic processing and stimulated the design of portable and compact equipment such as exploration needs. It is too early to describe specific accomplishments, but it can be predicted that compact long-period seismographs with digital recording will soon be developed along with small computers that will analyze seismograms for the arrival time, duration, envelope, and power spectrum of the separate wave trains.

Geothermal Transfer

A scientific problem of interest to exploration is the transfer of heat from the earth's interior to its surface. Birch (1954) suggested a relation between geothermal patterns and the oil productivity of geologic provinces. This hypothesis has not been examined by workers in exploration although more information is now available (Bullard and others, 1956; Lubimova, 1958) on thermal gradients and the rate of heat conduction in the earth. Swartz (1958) measured temperature profiles in deep holes on

atolls in the Marshall Islands and showed that there is a marked difference between the island profiles and the typical continental profiles.

Gravity and Geodesy

Scientific and applied geophysics are also closely related in the field of gravity and geodesy. The relation of gravity to the physics of the earth is reviewed by Garland (1956). Instruments for measuring gravity were first invented for scientific purposes, later made portable and accurate to serve exploration, and finally developed to use on ships and airplanes for making geodetic measurements. This permits the planning of a world gravity network for accurate measurement of the shape of the earth. The problem is classical, but it remained academic for a century after Stokes (1849) had set up the theory because there was no way of observing gravity over enough of the earth so that the problem could be solved. Bomford (1960) describes the present state of the problem; Hirvonen (1960) gives the most recent theory, and Heiskanen (1960) and Musen (1960) discuss the contribution made by observing satellite orbits. The present world network of absolute gravity (Woollard, 1958) is based on a series of pendulum measurements by Kühnen and Fürtwangler (1906) at Potsdam. Pendulum measurements are subject to systematic errors, and the Potsdam gravity was suspected, through measurements made by exploration instruments, to be in error by about 10^{-5} times the total. Preston-Thomas and others (1960) measured absolute gravity at Ottawa with a free-fall device with a probable error of 1.5×10^{-6} of the total. The results show the Potsdam value to be too high by 1.4×10^{-5}. The world network can now be corrected to the new absolute value and can be improved generally by enabling gravity meters to be more accurately calibrated.

Geology

The methods of exploration can be applied by the geologist to general scientific problems and will yield important information not otherwise available. Every advance in geology makes geophysics a more necessary tool for it. Thus exploration geophysics with its instruments developed only as prospecting devices may be the forerunner of the planet-scale geology of the future. A simple example of this is mapping the depth and structure of sedimentary basins. Mabey (1956) applied a combination gravity, refraction, and reflection survey to the intermontane basins in Southern California to determine the configuration of the bedrock and the thickness of the sedimentary fill. Johnson and Cook (1957) made regional gravity surveys in western Utah to discover fault zones and to compute the displacement of faults; they deduced from their observations the presence of pre-Basin and Range faults as well as the faults governing present-day structure. Pakiser and others (1960) used seismic, gravity, and aeromagnetic data to postulate that the Mono Basin in California was caused by extrusion of magma from a chamber at depth and subsequent subsidence along faults. Byerly and Joesting (1959) analyzed the salt anticline region of Colorado and Utah by means of gravity and magnetic data.

Miller and Ewing (1956) used a magnetometer towed by a ship in the Gulf of Mexico to make a total-intensity map of the area. They deduced, among other things, that the crust in the Gulf was of a thin oceanic type, and that the scarps bounding the calcareous banks were not tectonic. Berg and Wasson (1960) showed how the shape of buried thrust plates of the Wind River Mountains in Wyoming, and the configuration of sediments underneath them, can be found from seismic data. Allen and Smith (1953) applied gravity and seismic methods for measuring the thickness of the Malaspina glacier in Alaska.

The variations in thickness of the earth's crust are an example of a geologic problem that has so far yielded only to geophysical methods (Fig. 9). The question is important enough so that a plan is on foot to drill to it for samples that can be treated with the methods of subsurface geology. At present seismology has given the only direct solutions of the problem, although inferred solutions have come from gravity. Richards and Walker (1959) measured the depth to the Mohorovičić discontinuity in western Canada. They used regular seismic prospecting equipment and recorded at distances up to several hundred miles so the refracted mantle head wave would arrive first on the record. Shor (1955) observed reflections apparently from the top of the mantle. Widess and Taylor (1959) reported clear and persistent reflections from lower crustal layers in the Wichita Mountains in Oklahoma. They concluded that "the basement may exhibit all the seismic ingredients (acoustic contrasts, gentle dips and continuity) generally associated with sedimentary sections."

The inferential connection between gravity

and crustal thickness has been much studied because gravity observations are easier to make than deep refraction or even reflection shots. Woollard and others (1960) reported a study of crustal thickness through gravity, and Hales and Gough (1959) published a study of the same problem in South Africa. Gravity is also related to the depth of isostatic compensation. Ivanhoe (1957) gives an example of this by describing the gravity maximum in the Great Valley of California due to the isostatic effect of the Sierra Nevada. In continental geology oceans do, although on a smaller scale. The only practical way to measure this response is to observe the variations of gravity at a station while the earth rotates. The expected changes in gravity can be readily calculated. The discrepancy between the expected and observed change is due to the earth tide. Baars (1953) reported on a world-wide experiment on earth tides conducted with regular gravity meters. The subject has since been expanded into the study of earth vibrations generally; Benioff and others (1959) reported on the detection of

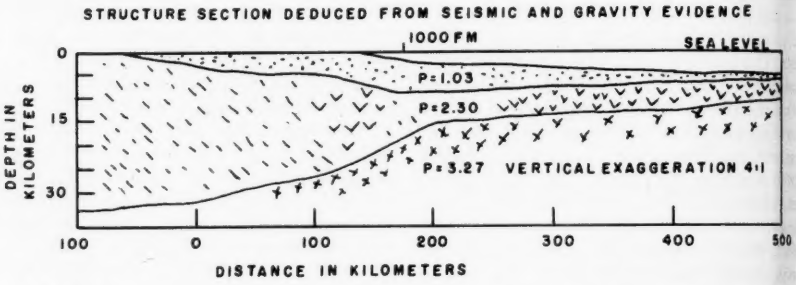

Figure 9. Crustal section extending from continent to ocean off Cape May, New Jersey (After Worzel and Shurbet, 1955, Fig. 6)

it is not known how long an area depressed into the mantle by the weight of mountains or glaciers will stay down after the weight is removed. The Fennoscandian shield (Hela, 1953) is known to be rising rapidly since the ice sheet melted. Innes (1957), however, shows that in northern Quebec a belt of strong negative isostatic anomalies seems to be due to granite masses dating from the Precambrian. This indicates local strength of the crust contrary to previous inferences about the mechanism of isostasy.

A similar problem is the residual effects of compaction on sediments when the weight is removed by erosion of the upper part of the column. It is not even known what the effect of compaction is, in the first place, on density, velocity, and heat conductivity, far less the relaxation after compaction stops. Parasnis (1960) discussed this problem and listed other attacks on it, but it is far from solved.

An important application of exploration instruments to problems in earth science is the study of earth tides. The rotating earth itself responds to the gravitational attraction of the sun and moon in the same way as the oscillations with periods other than the tidal periods. Clarkson and LaCoste (1957) described a gravity meter adapted to recording earth tides automatically.

RELATION TO INDUSTRY

Exploration geophysics receives economic support in two ways. A few oil companies maintain laboratories and field parties of their own and have continuous research programs. Most oil companies, however, employ consultants who operate field crews on contracts. The larger consulting firms have laboratories of their own. Significant advances and inventions have come from consultants as often as from oil-company laboratories so that much of the research and development in instruments is supported on a pay-as-you-go basis. Field work, on the other hand, is so expensive that consulting firms rarely do research in the field except on a contract. Field work for the purpose of extending knowledge instead of acquiring a piece of local information has thus to be financed by the production department of the oil company; this is not customary even when a large long-term saving might result. However,

much knowledge can be obtained by intelligent planning. Blundun (1959) described a campaign in Alberta planned so as to make sure nothing in the area was missed. Agnich and Dunlap (1959) attempted to define and codify the steps that are taken in a prospecting campaign, introducing the idea of a "performance standard" as a code of self criticism in an exploration program. Their position is that such a program lacks the self-checking devices that are built into science, so they propose to adopt for exploration the self-checking devices that modern business theory has invented for commercial procedures.

A good way to show how progress in exploration is affected by industry is to consider case histories. Nettleton's (1948) and Lyons' (1956) classical collections have been mentioned, and Peters' (1960) history of the Horse Creek field in Wyoming is a recent example. Appeal to case histories is one of the few ways in which geophysicists can check the rightness or wrongness of an exploration procedure. Unfortunately the method suffers from an almost irremediable defect—only the successes are published. Even if the failures were published they would not usually tell the geophysicist what went wrong because the data are not available; oil companies do not usually follow up a dry hole with the pattern of offsets necessary to get the information.

Applying geophysical techniques to the discovery of ore bodies is very different from the search for petroleum. The rewards are usually much smaller, and the methods cannot be codified so as to be usable by technicians: each problem is unique. Slichter (1955a) and Ellis and Blackwell (1959) offered plans for prospecting campaigns in mining, but published records show much greater diversity in problems attacked by mining prospectors than in those of searching for oil. Schmidt (1959) and Reichenbach and Schmidt (1959) describe, for example, seismic work around mines—a very special application. The recent collection of mining applications by the European Association of Exploratory Geophysicists (1958) is an excellent survey of modern mining geophysics.

A unique feature of the search for petroleum is that none of the ordinary methods of geophysical exploration looks directly for the oil itself. The search is for conditions associated with accumulations of oil. Except for geochemistry, which is not well understood and outside the scope of this review (*see* Ransone, 1958), no direct method has been found. A continual problem, however, is caused by the psychological lure of unorthodox methods of finding oil. Cook (1959) divides these methods into scientific, pseudo-scientific, and magical, with examples of each. These are of interest because laymen often charge technical people with having closed minds when they dismiss miraculous reports as impossible, and because scientists know that the true merit of an announcement cannot always be judged by the manner in which it is made.

SEISMOLOGY IN EXPLORATION

General Statement

We have seen that seismology is by far the largest division of exploration geophysics and the one in which there is most research effort. Because of the close relation between scientific seismology and seismology in exploration, advances made in theory in the former and in instrumentation in the latter tend to serve the whole field. These facts will now be elaborated, and some samples will be given of recent advances in seismological research.

Wave Theory

Wave behavior. The theoretical basis for interpreting seismological data, particularly in exploration, has always been the assumption that seismic energy travels according to the ordinary-ray theory of geometric optics. That is to say, once the velocities in the transmission medium are known as functions of position, the position of the wave front of the advancing energy at any time can be computed by assuming that the energy travels along rays whose direction is determined by Snell's law of refraction. To date no one has suggested that any other method of interpreting seismic data be introduced into exploration practice. However, Tolstoy (1959) recently pointed out that the theoretical validity of the ray theory in the case of seismic energy is sharply restricted by the requirement that the velocity (or at least a certain function of it) change very slowly with depth. This condition is violated by the fact that in well-stratified media, best adapted to refraction studies, there are actually discontinuities in the velocity. Tolstoy suggests, incidentally, that this may be the cause of the systematic differences between velocities as measured by refraction and by well logs. The thesis of his paper is that the rigorous solution of the problem of seismic energy in finite media is the normal-mode solution in which energy

travels, not in a wave front defined by rays, but as a result of the combination of the various normal modes of vibration of the block of material it is in. A simple example of this will be seen later in the case of seismic ringing or reverberation where a normal mode of a shallow layer of water sets up a continuous vibration.

A field experiment as a simple illustration of the normal-mode phenomena was reported by Press and Dobrin (1956). They examined the behavior of seismic energy traveling in a high-speed layer (Austin chalk, 95 feet thick) over a low-speed layer (Eagle Ford shale and Woodbine sand, 400 feet thick) and concluded that the high-speed surface layer acted as a high-pass filter for energy transmitted horizontally and as a low-pass filter for energy transmitted vertically. This is an elementary example of how modes of vibration affect the transmission of seismic waves and give a result which was beyond the scope of prediction by the ray theory.

The argument about modes and rays is carried a step further by Spencer (1960) who predicts the surface response of a stratified earth to a disturbance in which the assumptions of neither ray theory nor normal mode theory are valid. The treatment is theoretical and serves as an illustration of the use of the methods of Cagniard (1939). The same physical problem, that of the surface amplitude of reflected waves, is dealt with according to classical methods by Knopoff and others (1957) and by Tsepelev (1959), but the experiments to test their conclusions are not available.

Another theoretical attack on seismic problems, attractive because of its elegance and generality, is the application of reciprocity or symmetry developed for potential theory. If reciprocity can be proved to exist in a physical medium and a disturbance at one point gives certain effects at another, then the disturbance at the second point gives the same effects at the first point. This is a way of solving problems in wave behavior backwards, so to speak, when they cannot be solved forwards. Knopoff and Gangi (1959) apply this in the laboratory to some interesting model problems, with results that agree well with theory. White (1960) goes further and solves some idealized problems in radiation patterns that would otherwise be extremely difficult of solution; for one case he also performs a model experiment and confirms his theoretical results.

Wave shape. The next logical step in theoretical seismology after wave behavior—that is, the general theory of how waves are propagated in the earth—is to investigate what happens to their shape as they are transmitted. Transmission theory as ordinarily understood in, for instance, electrical engineering involves the solution of steady-state problems, such as what happens when a continuous pulse is sent through a transmitting medium. Seismology, on the other hand, is concerned with what happens to single pulses such as one emitted by an explosion or, in a more complicated form, by an earthquake. Ricker (1940) wrote the classical paper on pulse shapes. He stated four conclusions (p. 366):

> "1) a sharp seismic impulse gives rise to a traveling *wavelet* of shape determined by the nature of the earth's absorption spectrum, 2) a seismogram is composed of a succession of these *wavelets*, 3) the center of the *wavelet* travels with a velocity characteristic of the medium and 4) that the breadth of a given *wavelet* increases with increasing propagation time according to a definite law".

Much that seemed incomprehensible and illogical in seismogram interpretation was resolved by this approach, notably the supposed "character" of reflections caused by overlapping wavelets, the unstable patterns resulting from broadening, and the discrepancies in correlation resulting from following a single "leg" of a wavelet. The importance of the wavelet viewpoint for practical seismology was effectively summarized by Anstey (1956) in an article that epitomizes the whole business of obtaining seismograph records in the field and the factors that affect it (Fig. 10). The same author (1958) introduced a symposium on the subject of pulse shape in a paper which suggested that a realistic objective for pulse-shape study might be to obtain "from each geophone location a log of reflection coefficient against travel-time . . . free . . . from surface waves, transverse pulses, and multiple reflections." The author says four steps are necessary to reach this objective, the most important one being to apply an inverse wavelet operator which will turn the shape of the wavelet back to what it was when it was first reflected or propagated.

The problem of unscrambling seismograms has been attacked by solving it backwards—that is, by predicting what would occur if a given set of reflecting layers with given characteristics were present. Peterson and others (1955) wrote a classical paper showing how information from well logs could be used to

construct the seismogram that would result from shooting a reflection spread at that location. This permits a more certain interpretation of the record than was previously possible and enables the interpretation to be extrapolated from that location outward. Berryman and others (1958) extended the theoretical treatment and developed a formula that could be used in a digital computer. Wuenschel (1960) built on the previous work and described a technique to include multiples and transmission coefficients in synthesizing seismograms for programming in the IBM 704 computer. While, as the author says, "the mathematical model on which the computation is based is not as yet adequately realistic" the attack on the problem of multiple reflections, one of the most troublesome disturbances, seems to hold promise.

Practical seismologists have long known that the weathered (near-surface) layer has an influence on the shape of seismogram pulses out of proportion to its relatively small thickness. This was discussed by Gutenberg (1956) for earthquakes and is dealt with by Menzel and Rosenbach (1958), who explain the observed phenomena by a velocity change; they do not account for losses.

The related subject of instrumental distortion is examined by Landisman and others (1959), who show how the various kinds of instrumental distortion can be explained by the physical constants of the recording instruments.

Wave attenuation. Of great practical concern to the seismologist are the laws governing the strength or amplitude of the signals he receives. It is easy to see that the amplitude of a reflected pulse arriving at the surface will be partly determined by the nature of the surface at which it was reflected—that is, by the contrast in elastic characteristics between the material above the reflecting surface and that below it. The relative amplitudes of such pulses are therefore a clue to the physical character of the geologic contact between the upper and lower layers. Muskat and Meres (1940a) presented examples of the effect of different velocities and angles of incidence, based on a classical theory (Muskat and Meres, 1940b) which the authors applied. The subject was expanded by Berryman and others (1958), who considered the effects of multiple reflection and supported their results, qualitatively at least, with experimental data. It is unfortunate that the exploration industry long neglected the possibilities put within its reach by the study of differential amplitudes. The reason is that it was deemed more effective in field work to have records on which reflections had the same amplitude from the beginning to the end of the record, so that an automatic gain control was used. Some such device was necessary, but to make records look good it was applied with a short constant, so the effect of amplitude contrast was lost.

An application of amplitude study, which is perhaps more serious than the interpretation

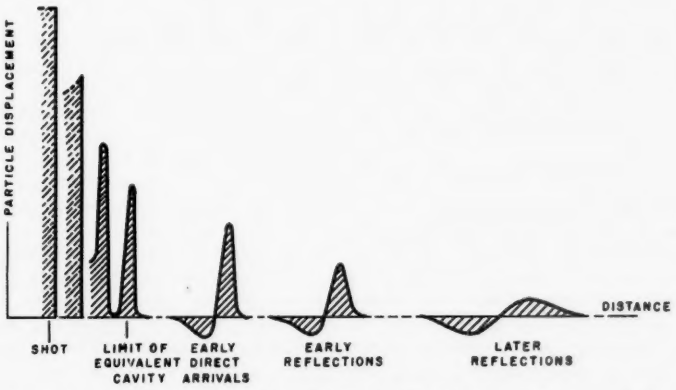

Figure 10. Qualitative illustration of change of displacement pulse form during propagation through the earth (After Anstey, 1957, Fig. 1)

of amplitude contrast, is the study of the energy lost by a seismic signal as it travels through different kinds of rocks. In the elementary theory of the transmission of seismic waves, it is assumed that the materials through which energy passes are perfectly elastic for "small" deformations; that is, that when stress is imposed on them the resulting strain is proportional to the stress, and that when the stress is removed the material reverts to its original state. This assumption implies that no energy is lost. Since attenuation or energy loss is actually observed, not to speak of the fact that physical intuition will not permit the use of a perpetual-motion model any longer than necessary, the description of energy loss is naturally used as a guide to the formulation of a new equation for wave motion in solids. Born (1941) considered the energy losses due to viscosity and solid friction and concluded from theory and experimental data that solid friction was the more important, which implies that the attenuation of energy in the earth is proportional to the frequency. Ricker (1940; 1953) presented mathematical solutions, with experimental studies to illustrate them based in the later paper on an extra term in the wave equation in the form of a time derivative of the strain to take care of imperfect elasticity. His results indicated that viscosity was predominant and that the attenuation was proportional to the square of the frequency. Collins and Lee (1956) added more terms to the resulting solutions, but the results were inconsistent. Busby and Richardson (1957) describe an ingenious method of measuring the attenuation of energy in sedimentary rocks as a function of the frequency but do not apply their results to theory.

The theory is again examined by Knopoff and MacDonald (1958) who state that "the experimental data require that deviations from perfectly elastic behavior cannot be described by any linear equation with constant coefficients" and that Hooke's law (strain proportional to stress) must fail for "extremely small strains." An experimental determination of the attenuation in a thick layer of Pierre shale in Colorado was made by McDonald and others (1958) who concluded that the shale did not behave as a visco-elastic material conforming to Ricker's (1953) theory. Horton (1959) presents a table demonstrating the various theories of wave behavior in the form of equivalent mechanical and electrical circuits, and proposes one of these with a second-time derivative term which gives the results observed in the case of shale. Peselnick and Zietz (1959) measured the attenuation in fine-grained limestones at ultrasonic frequencies and concluded that the solid-friction type of attenuation was predominant.

The essence of the physical problem is that when the material through which the energy passes is deformed it does not return to its original state, and the search for a wave equation that will predict the observed results is primarily a search for expressing that fact in mathematical language. Collins (1960) makes a further attack on the problem but concludes that it may be necessary to compute the results given by spherical waves (so far plane waves have been used) to get a satisfactory agreement between theory and observation.

The problems of wave shape and wave attenuation have been discussed under different headings because of the tendency of investigators to put observations of the two characteristics in different categories. Energy losses can be measured quite effectively in the laboratory, while pulse shapes have to be observed through the behavior of a model or else in the field with apparatus that tends to distort them. The problems, however, are not in fact separate problems. If we knew what happened to energy in rocks—if we were able to predict the displacement of particles as a function of time and the physical characteristics of the rocks—we should have solved the problem of pulse shape as well as that of attenuation.

Wave scattering. An interesting part of the theory of elastic waves in the earth is the study of scattering. As yet no application of the theory of scattering has been made to the practice of oil prospecting. However, Stilke (1959), after developing the theory of scattering for the case of an infinite horizontal cylindrical hole, suggested that it might explain certain irregularities in the seismic records in the neighborhood of a mine. It seems reasonable to predict that, when more experimental work is done on the subject, it will be found that faulting produces scattering which can be noted on seismic records and used as a diagnostic criterion. Rieber (1937) long ago showed in a model how a faultlike structure would act as a new source, but he did not discuss the theory of the phenomenon. Recently Knopoff (1959) presented the theory of scattering of elastic waves by a spherical obstacle, and Miles (1960) presented a similar study for the effect of small inhomogeneities. As would be expected, the results in each case depend strongly on whether

the disturbing body is large or small compared to the length of the seismic wave that it disturbs.

Shear waves. Elementary wave theory shows that, when an explosion or other disturbance occurs in a liquid, the resulting energy radiates outward from the source in the form of compressional or longitudinal waves. When a compressional wave travels through a transmitting medium, the individual particles move in a line along the direction of propagation, with no component of sidewise motion. The determined for the purpose of computing (with the aid of the density and the compression-wave velocity) the elastic constants of the transmitting medium. Ricker and Lynn (1950) observed them with exploration equipment modified for the purpose and applied them to a particular case. As far as the reviewer knows, this lead has not been followed in practice.

If shear waves could be used for exploration, several advantages might result. For instance, Jolly (1956) pointed out that seismograms of horizontally polarized shear (*SH*) waves are not

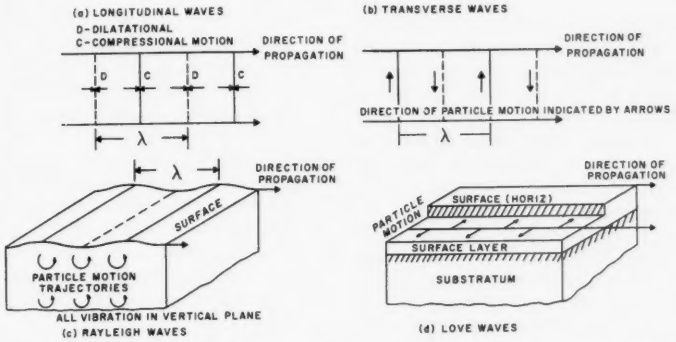

Figure 11. Characteristics of elastic waves in solids (After Dobrin, 1960, Figs. 2–3)

other type of wave, the shear or transverse wave, causes the individual particles of the transmitting medium to move in a direction crosswise to the direction in which the wave is traveling (Fig. 11). Shear waves are not transmitted in liquids because liquids have no rigidity. That is to say, moving particles in a liquid have so little tendency to drag other particles with them that shear waves do not propagate.

In a medium that has rigidity shear waves can of course be propagated. Earthquakes, which apparently always include a wrenching or sidewise motion, send out shear waves abundantly. Explosions, being phenomena of pure expansion rather than slipping, are supposed to send out compressional waves only—at least to begin with. It is for this reason that shear waves have not been used in the regular practice of exploration. Compressional waves, however, are partly converted to shear waves when they are obliquely reflected. This causes some shear waves to be generated by explosions. Such waves can be recorded by directional seismometers, and their velocities can be transformed in kind when they are reflected or refracted from a horizontal surface, as are compression (*P*) waves. The chief difficulty with using shear waves in exploration is that the methods reported by Jolly and by White and others (1956) of producing shear energy are insufficiently powerful although they produced waves that were qualitatively satisfactory. Volin and Rudakov (1956) reported that better results could be obtained by directed explosions. Evans (1959) reported good agreement with theory in experiments with models. He concluded that *SH* waves produced simple and clear reflections and that they could be generated in an elastic solid but he found that a thin low-velocity layer gives rise to Love waves on the surface of such amplitudes as to interfere substantially with the otherwise clear *SH* reflections. (Love waves are *surface* shear waves, horizontally polarized, Fig. 11.)

Wave study with models. One of the most powerful tools in seismology is the method of models. An important disadvantage of exploration geophysics is its inability in general to perform experiments. Even when the interpre-

tation of seismic data is checked by drilling, the information obtained is usually sketchy and ambiguous. Most interpretations, of course, are never proved either right or wrong by the drill, and the complicated reshooting procedures that would be necessary to demonstrate that an interpretation is at least qualitatively correct are almost never carried out. Model studies constitute an excellent if incomplete method of checking some of the conclusions of seismologic theory.

Terada and Tsuboi (1927) studied elastic-wave propagation by sending sound at ultrasonic frequencies through small-scale three-dimensional models. Such work is cumbersome because it is difficult to find materials with suitable elastic constants for model use and difficult to handle them when they are found. A simplification results if the number of dimensions is reduced from three to two. By suitable modifications of the theory, materials in the form of thin discs may be used to simulate the transmission of sound in earth, and experimental solutions found for many theoretical problems. Oliver and others (1954) gave the theory and laboratory details for a set of experiments with thin discs of metal, plastic, and even paper. The same authors (Press and others, 1954) published a second paper using their methods for two- and three-layer refraction problems (Fig. 12). They encountered the same discrepancies between observations and elementary theory as are found in full-scale field work. In both papers the relation between wave length and the thickness of the layers is shown to be a decisive factor in determining the quality of the results.

The model method has been studied diligently since 1954. Hall (1956) reverted to a three-dimensional model and observed reflections from the bottom of a tank full of water and from a horizontal slab of slate in the bottom of the tank. This arrangement permitted a number of geometric configurations and in consequence a good opportunity to study how the pulse shape varied as it traveled greater distances and was reflected from different interfaces. Carabelli and Folicaldi (1957) carried the process a step or two further and observed reflections from layers whose thickness was much less than the wave length of the reflected energy. The variation in amplitude of the reflected wave as a function of the thickness and the sound velocity of the reflecting layer was studied. The authors discovered that very thin layers—one-fiftieth of the wave length in thickness—, not differing greatly from the surrounding material in elastic characteristics, give reflections of unexpected strength. De Bremaecker (1958) studied the behavior of a surface wave incident on a corner. Riznichenko and Shamina (1959) reported further studies with three-layer models, with variations in the thickness of the layers. Angona (1960) introduced simulated faults and curved surfaces into the model and showed the effects of diffraction (noticed by Rieber, 1937) and the contrast between the behavior of diffractions and reflections.

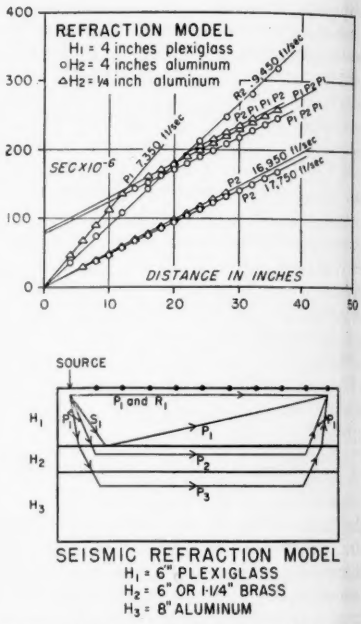

Figure 12. Seismic-refraction model showing paths of pressure (*P*) waves, shear (*S*) waves, and Rayleigh (*R*) waves (After Press and others, 1954, Fig. 8)

Extracting Signals from Noise

Importance of signal enhancement. So far we have been dealing with questions that are generally lumped into the category of wave theory. Except perhaps for the topic of scattering, the subject is the behavior of seismic energy as a whole in its course through one or more transmitting media. The importance of this subject for the future of seismology needs

no emphasis. Advanced wave theory, however, has not affected the general practice of exploration seismology at the field-computing level. Routine interpreters of seismic data use no wave theory more complicated than Snell's law for their regular interpretation work. The researches reported here of wave theory are still academic as far as daily work goes. When it comes to extracting the signal from the noise, however, a problem of immediate practical significance appears. Although the routine seismologist may be indifferent about the theory of wave transmission, his attention is closely engaged by the question of noise. This is doubly true because every current prospect is in a sense a marginal or frontier problem. What makes a seismic exploration prospect marginal is always the difficulty of finding reflection events on the records. If reflections can be found readily the records are said to be noise free or relatively so. When reflections cannot be seen readily it is assumed that they are actually or potentially present but are obscured by the noise. It is natural then that the most important practical work done by laboratories and by theoretical workers is to devise methods of making the reflections visible through the noise.

Seismic signals are extracted from noise in different ways. The simplest is to place the seismometer array far enough from the shot point so that the reflections, traveling in a high-velocity medium, arrive before the disturbing surface waves. A refinement of this method is to dispose them half a wave length apart with respect to the surface waves so that these waves tend to cancel themselves. This is the method of multiple seismometers. A somewhat more refined way to get rid of noise is to use electrical filters. This works when the noise frequency is substantially different from the frequency of the reflections. If the noise frequency is nearly the same as the signal frequency, more complicated methods of signal enhancement must be used.

Multiple seismometers. The method of minimizing noise by using arrays of seismometers instead of individual units so that certain types of noise tend to cancel each other has been known and practiced since the earliest days of reflection shooting, and the first volume of GEOPHYSICS contained an article by Klipsch (1936) in which he evaluated its advantages by statistics. Many papers dealing with the subject have appeared since then, among them one by Hales and Edwards (1955) applying antenna theory and showing how the response of an array depends on the direction of incidence of the wave and its frequency. The problem can be treated more powerfully if it is approached by studying the noise the array is supposed to reduce, as well as the signal it is supposed to enhance. Horton (1955; 1957) presented an analytical description of the noise, which he

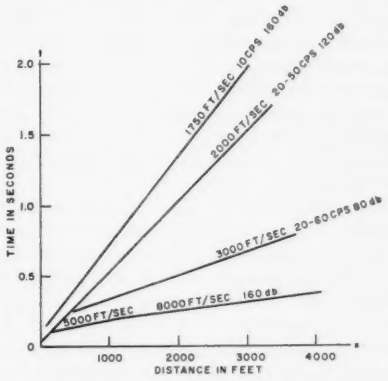

Figure 13. Time-distance plot of noise waves at a recording location (After Smith, 1956, Fig. 5)

said was due to numbers of "scatterers" in the earth through which the reflections travel. The scatterers are

"caused by local variations in the physical properties of the earth, by bedding planes of limited horizontal extent, or by any other type of departure from homogeneity". . . . "If it is assumed that the spatial distribution of the scatterers is random, but that the shape of each reflected pulse is the same, then one may deduce the statistical properties of the seismogram trace from the statistical theory of noise."

Smith (1956) made a comprehensive study of the theory and application of multiple arrays, showing a specific example of a set of individual noise waves at a given location (Fig. 13). He applied generalized harmonic analysis (Wiener, 1933) to the problem and did not make restrictive assumptions such as those of a plane-wave signal and random distribution of noise. The result was a set of conclusions determining the optimum recording system, based on the best filter setting, seismometer array, shot array, shot depth, and offset distances.

Shot arrays have not previously been men-

tioned, but they are an important part of exploration practice and are logically included in the topic of multiple seismometer arrays. They were first described by Poulter (1950) who used multiple charges in the air and on the surface in difficult areas. A more recent article on patterns was published by Muir and Hales (1955) which showed clear examples of the efficacy of pattern shooting.

A full-scale, rationally conceived exploration problem embodying the available techniques for analyzing noise and setting up the shot and seismometer arrays in accordance with the findings is described by Graebner (1960). He shows how an area in which the records were previously too poor to use was successfully explored by making a noise study for the particular area.

Filtering. The part of a seismic-exploration system whose design and use are the most critical to successful practice is the filtering system. The reason is that the undesirable signals, such as the surface waves resulting from the explosion, are frequently so strong as to make the desirable signals, such as reflections, practically invisible in a true recording of the motion of the earth following a shot. The regular way to solve this problem is to exclude the surface waves ("ground roll") with a filter that cuts out the low frequencies and to exclude such disturbances as wind noise with a filter that cuts out the high frequencies. That solution has the defect that the residual signal—what is left over—is necessarily distorted. In particular it is lengthened, so that a pulse originally 0.02 second long may acquire a "tail," which effectively doubles its length (Fig. 14). In addition to lengthening the pulse, the filter shifts the phase—that is, it displaces some characteristic point of the pulse, such as its first peak, several thousandths of a second on the record. Filtering should, therefore, be done with restraint in the field, especially since there is a tendency to use filters to make the records look smooth and regular. There seems to be a gap between the theory of filtering, which has been well carried forward, and the knowledge of how to use filters in the field so as to extract the most geologic information from the records.

Distortion of reflections due to instrument design, especially the filters, was discussed by Anstey (1956) who described the character and time distortion in the filters and in the automatic gain control, and harmonic distortion (the addition of signals whose frequencies are multiples of the original signals) as introduced by the amplifying system. The paper is an excellent guide to the elementary practice of field seismology. A theoretical study of the distorting effect of filters was presented by Holtzmann (1959) showing how the true time of arrival and length of the arriving pulse could be computed with the aid of the power spectrum of the pulse. An application of filters to

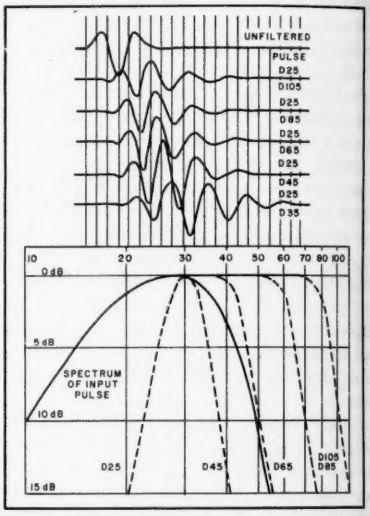

Figure 14. Effect of filtering with different band widths (After Anstey, 1956, Fig. 2)

the problem of getting rid of multiple reflections, based on the assumption that a primary reflection has a higher frequency than its multiples, is described by Epinat'eva and Ivanova (1959); their demonstration is qualitative and is not shown to be general. A more general application of frequency filtering is suggested by Klushin (1959) who shows how geophysical anomalies can be lifted out of the "mean square error" if the frequency spectrum of the anomaly differs from that of the error.

So far we have been dealing chiefly with ordinary electrical filtering—that is, putting the signal through a resistance-capacitance network and changing its character by the methods used for steady-state alternating current. Recently, a new kind of filtering has been developed and put into use; it constitutes a much more powerful tool for seismology than the elementary techniques of arrays and electrical

filtering. The new filtering techniques constitute "operating on" the data rather than merely putting the signals through an electrical circuit. An example of operating on seismic data was described by Ricker (1953) who put the data through a "wavelet contractor." The purpose of the wavelet contractor was to reverse or *equalize* (as in communication theory) the effect of the transmission medium and concontract or narrow the wavelet to make up for the broadening it underwent in transmission. The advantages in making seismograms more readable are obvious. In the same year Wadsworth and others (1953) described the application of linear operators to seismic data. A linear operator is a series of fractions, each of which is assigned to a time in a regular sequence of times preceding and following a certain zero time. For each point in time on the record, the amplitudes of the trace corresponding to the sequence of times before and after the point are multiplied by the fractions, and the whole added to form a new value for the filtered trace. The process is then repeated for the next index point on the record, say 0.002 or 0.005 second later. Holloway (1958) has described the theory of it clearly. Such an operation must be done on a computing machine since so many individual multiplications and additions are involved in operating on even a single trace. However, once the appropriate machine is available and the seismic records are in suitable form, all kinds of ordinary filtering, plus the newer kinds, can be done quickly and with a greatly increased usefulness to the exploration seismologist.

A comprehensive review of the subject of filtering was published by Smith (1958). He points out that whereas the conventional filters could fairly well be characterized by giving their cut-off points in cycles per second and their rejection rates in decibels per octave, the new filters cannot be described so simply. In particular he describes (1) digital filtering, (2) delay-line filtering, (3) cross-correlation filtering, and (4) zero-phase filtering. Delay-line filtering is a form of digital filtering; it is described by Jones and others (1955). Cross correlation is done with linear operators; it is described by Jones and Morrison (1954). An adequate description of these types of filtering is beyond the scope of this review.

Other applications of computer techniques are reported by Robinson (1957) who shows how wavelet shapes may be extracted from records, thus leaving their amplitudes as a residual, and by Lindsey (1960) who applied the autocorrelation function of the seismic trace to detecting ghost reflections. (A ghost reflection is a pulse reflected from the surface after an explosion and therefore following in reversed form the original pulse.) There is no reasonable doubt that more advanced and comprehensive applications of computing-machine techniques will soon appear.

Ringing. The most dramatic illustration published to date of the application of computer techniques is the solution of the problem known as "ringing." The process of operating on seismic data for the purpose of removing noise or for the purpose of retransforming it back to the shape it had before it underwent transmission through the earth is not easy to explain in nonmathematical terms, and no apt analogy, of the sort that can often be drawn to explain the solution of a physical problem, seems to present itself. Fortunately, the history of the problem of "ringing" is available to serve as an example of what the new techniques can do.

Ringing is encountered in reflection shooting in shallow bodies of water where the depth remains more or less constant at 100–200 feet. The water-to-bottom contact and the water-to-air contact are both good reflecting surfaces, which appears to result in the fact that when shooting is attempted the record traces look like either continuous sinusoidal waves or a regular sequence of constant-amplitude signals, instead of being a succession of pulses at arbitrary intervals that can reasonably be treated as reflected events (Fig. 15). This is ringing, and it was early recognized that the problem was probably one of reverberation. Burg and others (1951) described the problem clearly and showed that the observations could be explained by treating the water layer as a wave guide and studying the normal modes of vibration in it. While the description more or less accounted for the observation, no specific remedy was offered. Sarrafian (1956) used a model in which he found the behavior of the pulse shapes and frequencies to be similar to that encountered in full-scale shooting, and he observed multiple reflections. He did not, however, suggest a practical solution to the problem. The nature of ringing, or "singing," was more closely inquired into by Werth and others (1959) who conducted field experiments and found that the description offered by Burg was not followed exactly by their observations. They reasoned that the phenomena they ob-

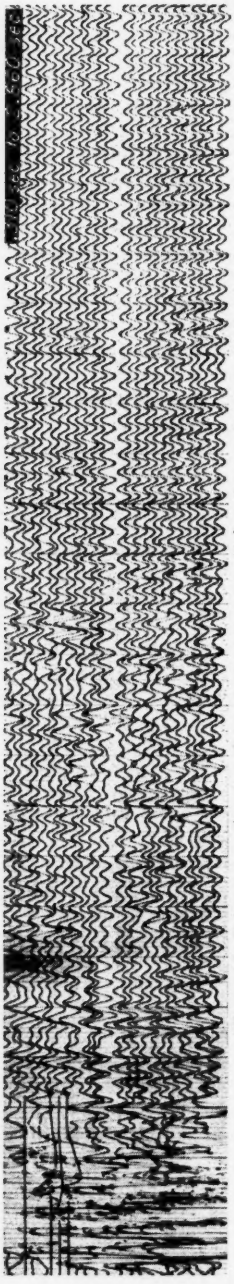

Figure 15. Seismogram showing almost pure sine waves (After Burg and others, 1951, Fig. 2)

served called for an "extended" source, outside the boundaries of the wave guide, and they concluded that the extended source was in reflections from deep layers that reverberated in the water after having been reflected and returned to the surface. They caution the reader that they have seen "ringing" records to which their theory does not apply and they suggest that ringing may sometimes be due to causes other than the one which they observed. Liu (1959) made a theoretical investigation of the problem, beginning with the classical pressure-wave equation, and solved it for a point source in the water layer. This provides an explanation for the way the frequency depends on depth and predicts the transient amplitude response. However, Liu is not satisfied with his model because it predicts that the energy will decay sooner than it actually does.

It remained for Backus (1959) to offer a solution for the problem. He treated the water layer as an extremely sharp linear filter and showed how its acting as an "energy trap" would produce the observed effects. He also showed various ray paths, including a deep reflection which was reflected up to three times in the energy trap (Fig. 16). Finally, by studying the observed data, he built up a transfer function as an expression of what the transmission path did to the energy as it passed through. The original pulse, operated on by the transfer function, gives the recorded signal. Then the recorded signal, operated on by the *inverse* of the transfer function, changes the recorded signal back to an approximation of the original signal (Fig. 17). This improves the record substantially by making visible some reflections imperceptible in the untreated record.

Seismic Velocities

Importance in geologic interpretation. So far in this review the discussion of seismology has dealt with problems that are to be solved by a combination of theoretical study and instrument construction and so are removed in some degree from the day-by-day practice of translating seismic data into geologic information. We now leave the basic questions of *why* seismic waves behave as they do and approach a set of topics better described as *how* seismic waves act and therefore more directly applicable to field practice.

The first of these topics is the velocity of sound in the earth. The importance of the velocity of seismic energy in the reduction of

seismic data to a geologic picture lies in the fact that the velocity varies, both with the depth and with the horizontal position. If the velocity were constant there would be no problem. The observed travel times (with minor corrections) could be plotted on a map or a cross section, and a distortion-free diagram of geologic structure would appear. Since velocities are not constant, the travel times must be converted into depths, or the geologic picture will be distorted. For instance, if the line rocks that occur in large masses, so that correct velocities can be computed from laboratory specimens of such rock. In seismic exploration with reflections, however, the energy travels through sedimentary rocks that are weathered, unconsolidated, stratified, inhomogeneous, and anisotropic. In consequence the velocity along a given path is not generally predictable from laboratory data and must be found in other ways. These ways can be divided roughly into measurements that can be per-

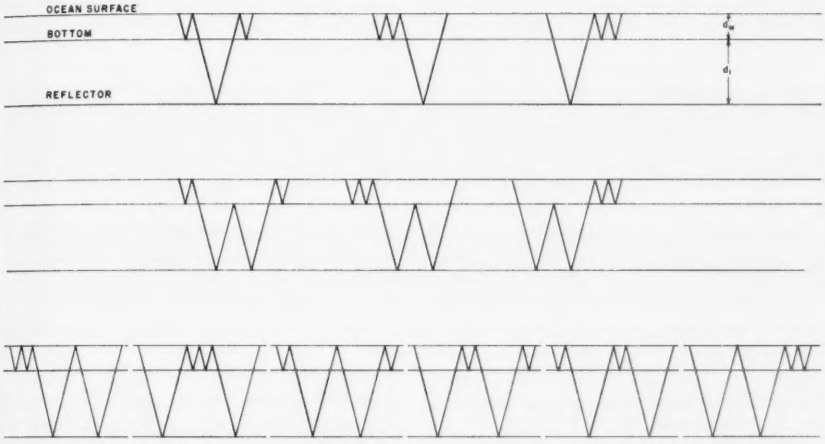

Figure 16. Ray paths showing multiple reflections in energy trap (After Backus, 1959, Fig. 6)

velocity increases with depth, the deeper formations will be mapped as thinner and shallower than they are unless the increase is taken into account. If the velocity varies laterally the consequences may be more serious as the structure, as mapped, can be either exaggerated or masked. Oil fields have been found by taking into account lateral variations that were ignored by earlier prospectors. Laubscher (1956) shows an example of how the interpretation of faulted structures is distorted by lateral variations in the velocity. It is thus necessary to know as much as possible about how the velocity varies in order to interpret seismic data correctly.

The velocity of sound in the earth is measured in various ways. The theory of seismic energy relates its velocity of propagation directly to the elastic constants of the medium through which it passes, provided the medium obeys certain idealized laws. This view of the problem is good enough, in the case of crystal-

formed on the surface and measurements requiring the use of bore holes for the detectors.

Surface methods of measuring velocity. The first methods to be used in measuring velocity were naturally those that could be carried out on the surface since laboratory methods had not been developed, and few bore holes were available into which measuring instruments could be lowered. Leet and Ewing (1932), for example, measured the velocity of granite by timing an impulse from a shot point to a recording point on a massive outcrop. The velocities of buried rocks were measured by refraction; the recording points were set far enough away so that the first-arrival energy traveled through a deeper and faster formation, thus overtaking the energy that traveled through the slower and shallower layers. This method had the disadvantage that the velocities measured were those of the high-velocity formations in the sedimentary column, and therefore not typical of the column as a whole. At

present in refraction work the observed velocities are corrected by an assumed factor, but the method is still an approximation. Nafe and Drake (1957) illustrated a modern use of the refraction method and discussed the relation between porosity and velocity in shallow- and deep-water sediments. Vertical velocities were measured from the surface by the method of velocity profiles as described by Green (1938). In this method the travel times to a given reflecting point are observed with the shot point close to the recording point and again with the shot point a distance from the recording point. This permits the velocity to be

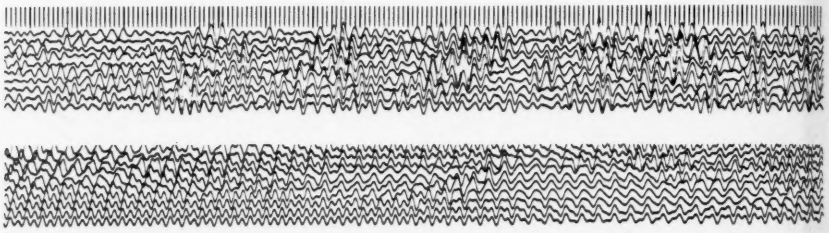

Figure 17. Record from Lake Maracaibo before and after inverse filtering (After Backus, 1959, Fig. 17)

calculated, but the method becomes inaccurate as the depth increases.

Laboratory measurements of velocity. Measuring seismic velocities of laboratory specimens of rocks required the development of special techniques for putting an energy source into one end of the specimen and measuring the elapsed travel time at the other. A method of doing this was described by Hughes and others (1949). Its application to the measurement of velocities of dunite, granite, limestone, sandstone, shale, and marble, at high temperatures and pressures, was reported by Hughes and Cross (1951). Baule (1953) used a magnetostrictive pulse for the same purpose. Recently Birch (1960) measured velocities at pressures up to 10 kilobars; his paper contains a good history and summary of the method and a useful bibliography. Knopoff (1954) gives the results of some laboratory determinations of velocity and suggests that they agree with field measurements (Leet and Ewing, 1932) better than previous laboratory results because the pressures used closed the cracks and flaws in the specimens. Murphy and others (1957) used the method on synthetic cores. Wyllie and others (1956; 1958) used a compaction apparatus to measure the velocities of dry and saturated aggregates and suggested an empirical formula relating the velocity to various characteristics of the aggregate. Kuiper and others (1959) used a resonance method to measure the elastic constants of different kinds of limestone and succeeded in obtaining results that were more self-consistent than those previously published. The theory of a resonance chamber for measuring the velocity and attenuation of sound in water-saturated sediments was presented by Toulis (1956), and some measurements with the chamber were described by Shumway (1956; 1960). Laughton (1957) measured the velocities of sediments with a pulse technique at high pressures and observed shear waves when the sediments were sufficiently compacted by pressure. He also studied anisotropy in the laboratory by observing that as samples were compressed the velocity in the direction parallel to the compression increased faster than the velocity in the direction of propagation.

All laboratory measurements of velocity, however, are of doubtful validity because they are made at high frequencies, and it is not known how well they can be extrapolated to the very low frequencies used in seismic prospecting.

The wealth of experimental work that is being done (the quoted references are only a selection of the published literature on the subject) demonstrates the importance of better knowledge of the velocity in unconsolidated as well as consolidated sediments. The problem of what velocities obtain at what depth, and how they vary laterally, at an arbitrary location, is not soluble by laboratory methods unless the character of the sediments is much better known than it usually is in practice. It follows that, for present exploration purposes, the

field measurements of over-all velocities are of more immediate importance than are those made in the laboratory.

Field measurements of velocity. For velocity measurements, exploration depends more on data from bore holes than from any other source. The essence of the method is simple: a seismometer is lowered into the bottom of a bore hole, a shot is set off in an ordinary shot hole close to the surface, and the travel time of the impulse to the seismometer is measured. The seismometer is then lifted a short distance from the bottom, and another travel time is measured. A direct measurement is thus obtained for the one-way velocity from surface to a given depth, over approximately the same path as a reflection would travel. Some questions remain as to whether the one-way travel time is really half the reflection time, because of energy traveling in the well casing and the seismometer suspension, but well-velocity surveys are still the basis of most of the velocity information used today.

Kokesh (1952) described the advantages of making well-velocity surveys with the explosion in the well and a seismometer array on the surface. Von Zur Mühlen and Tuchel (1953) gave a general review of well velocities in West Germany, with remarks on the apparent influence of lithology and depth. Wyrobek (1959) made a similar, perhaps more analytical, study of the well velocities in England. Both authors followed the lead of Faust (1951) who made a statistical study of the effect of age and depth for shales and sandstones, and evolved a general formula for over-all use when direct velocity information is not available. In a later paper Faust (1953) introduced a lithologic variable whose value should be deducible from an electrical log if such a log is available from a nearby well. Chereau and Ledoux (1959) made observations on how the existence of lateral variations could be predicted from geologic data, and Acheson (1959) suggested a method of correcting maps for such variations.

A new method of obtaining seismic velocities from wells was described by Vogel (1952) and by Summers and Broding (1952). The method is called continuous-velocity logging, and the principle of it is to measure the velocity over small intervals by lowering into a well a device that sends and receives an acoustic pulse across an interval of 5 feet or so. The method is subject to various corrections and inaccuracies. The errors and corrections are critically discussed by Schwaetzer (1958), and the low precision of the method is pointed out by Hicks (1959) and by Wood (1959), who showed a discrepancy between the continuous-velocity method and the seismometer method.

Continuous-velocity logs have nevertheless been applied to geologic problems, presumably on the theory that the local variations are significant even if the over-all results do not agree with standard seismometer measurements. Hicks and Berry (1956) studied the relation between water saturation and the continuous-log velocity, and Dunoyer de Segonzac and Laherrere (1959) used the continuous log to measure anisotropy in the Northern Sahara.

Anisotropy. The velocity parallel to the surfaces of stratification is different (larger in general) from the velocity normal to those surfaces. Cholet and Richard (1954) described a simple method of measuring anisotropy in the field by conventional bore-hole methods and defined an anisotropy factor for which they submitted several values. Postma (1955) offered a theory to account for anisotropy, and Uhrig and Van Melle (1955) presented some further measured values for the anisotropy constant. Kleyn (1956) discussed computation methods used for an anisotropic basin in Sumatra, and Helbig (1956) presented a model study with an anisotropic wave-front analysis.

Sources of Seismic Energy

Explosives. The first source of energy used by systematic seismology was of course the natural earthquake. Large earthquakes release energy that can be detected anywhere in the world and that actually sets the whole earth into vibration. While the science of seismology found its beginnings in deductions made from observations of earthquake waves, it has never been possible to utilize earthquakes for the smaller-scale investigations required by the exploration for minerals. To set in motion the waves that are observed by exploration seismologists, there is no energy source better, cheaper, or lighter than ordinary dynamite. Mechanical sources have been tried and will be discussed later, but none of these has had more than a limited application, chiefly because of the quantities of energy needed to achieve the necessary travel distance. The perceptibility of seismic energy at a receiving point is limited by the level of seismic noise in the earth. While modern techniques are capable of extracting a signal from noise when the signal has a much lower level than the noise, for practical purposes it is easier and cheaper to shoot a few

extra pounds of dynamite in the hole than to resort to signal-enhancement techniques. In practice this limitation has so far ruled out all energy sources except high explosives.

Considering the importance of the explosive charge to the day-by-day needs of exploration, little is known about the mechanics and energy transfer involved in a detonation. Sharpe (1942a) listed some well-known empirical facts: increased energy results from firing in sand or clay rather than limestone and from firing in an enlarged cavity; the frequency of the received pulse increases with velocity of the medium in which the charge is exploded, and decreases with the charge size; high-speed explosive (dynamite) is better than low-speed (black powder). He developed a theory of explosions based on a sphere at whose surface elastic laws began to be obeyed and he showed the importance of the radius of this "equivalent cavity" to the character of emitted energy. The same author later (Sharpe, 1942b) made observations of ground motion inside the equivalent cavity. Morris (1950) divided the explosion into three states: the formation of the pressure pulse, the crushing of the surrounding medium, and the formation of a spherically spreading elastic pulse after the pressures stop breaking up the medium. O'Brien (1960) presented a theory relating weight of explosive to amplitude of disturbance, based on the postulate that (p. 29) "the radiated pressure pulse begins to obey infinitesimal strain theory once its impulse per unit area has decreased below a limiting value. This value is constant for a given rock type." O'Brien concluded that, contrary to Sharpe's evidence, the frequency content ought not to be affected by charge weight. His results for the relation of amplitude to weight were borne out by experiment in the case of ground explosions but not under-water ones. He also postulated a state of nonlinear action between Morris' second and third stages. Howell and Budenstein (1955) made observations of ground motion near explosions and arrived at a description of the energy distribution among different kinds of waves.

On the practical side, much has been done by the exploration industry to improve the pulse content of seismic records by novel arrangements in detonating the explosives. Poulter (1950) described the advantages of detonating explosives in a pattern instead of in a concentrated charge and on the surface or even suspended above the surface instead of buried. Musgrave and others (1958) examined the directivity effect of charges that were elongated instead of concentrated.

Mechanical sources. Efforts have been made since the beginning of exploration to get energy into the ground by mechanical rather than explosive means. The object was to control the source better, to take advantage of a continuous source, or to propagate energy at higher frequencies than appeared in pulses from explosions. Howell and others (1940) reported experiments with an electrodynamic shaker and a magnetostriction tube, using frequencies of 400–1500 cycles. The investigation was continued by Evison (1956; 1957) who succeeded in making a closer approach to a geophysical application. Crawford and others (1960) recently reported an attempt to exploit the advantages of a continuous source with variable frequency. So far published reports of these methods do not promise an application to deep exploration because of their limited range, but the problem is currently under attack, and results are expected.

The one mechanical method that has been applied commercially is dropping a heavy weight on the ground a number of times and adding the resulting records electrically. A good description of the method is given by Neitzel (1958), and examples of records are given by Domenico (1958). Although the method is too cumbersome for use where dynamite gives good records, it has proved useful in areas where records of dynamite explosions were poor.

Interpretation

Elements. The interpretation of seismic data is the art of deducing a piece of geologic information from the record of an earthquake. In scientific seismology the most important data on the record are the relative arrival times, amplitudes, and frequencies of the many wave trains observed on the record. In exploration seismology, to date, the shear waves and surface waves are ignored. The only data used in interpretation are the arrival times of successive compressional waves assumed to be reflected or refracted from a geologic discontinuity, and the direction in which the energy is traveling when it reaches the recording point. The travel time can be known because the time of the explosion is electrically marked on the record; the direction of travel can be deduced from the difference in arrival time at the ends of the seismometer array. From these data, plus an assumption as to the velocity of compressional waves at each point through which the energy travels, a computation can

be made which gives the depth and attitude of whatever geologic contact reflected or refracted the pulse whose arrival is being studied. This description applies to all seismic interpretations being made today, with unimportant exceptions. The method is described by Nettleton (1940) and other authors of textbooks. Its theoretical basis is the ray theory of optics in which Snell's law of geometric optics determines the direction of the rays in which the energy travels. Energy in the so-called normal mode of vibration (*see* section on Ringing) is considered to be noise and must be largely eliminated from the record before it can be interpreted. Shear waves (*see* section on Shear Waves) have been studied but are not used in regular interpretation. Surface waves are also regarded as noise, although Press (1957) proposed to adapt for exploration purposes a method he uses to measure the thickness of the earth's crust. This is done by observing the dispersion—that is, the difference in velocity between the pulse as a whole and some recognizable part of the pulse such as a peak in its record trace. As far as the reviewer knows this suggestion has not been followed up.

Ray geometry. Since the observed arrival time, the observed arrival direction, and the velocity distribution in the earth are the only elements of the problem (except for surface corrections) considered today, it can be said that, whenever these elements are sufficiently well determined, the problem of interpretations can be solved. Measuring the first two of these elements is primarily an instrumental problem. The velocity distribution has to be assumed because it can never be fully measured.

The simplest assumption for the velocity is that it is constant. In this case the rays for energy travel are straight lines, and the interpretation geometry is elementary.

The assumption with one more degree of complexity is that the velocity varies with depth. In general, a linear relation ($v=a+bz$) or square root ($v=a(1+bz)^{\frac{1}{2}}$) (v=velocity, z=depth, a and b are constants) is a close enough approximation for all practical purposes. Slotnik (1936) described the ray geometry in the two cases and showed how to compute the depth and attitude of the reflecting plane. In cases where the velocity is thought to vary with depth in such a way that it cannot be represented as an integrable analytic function of depth, a solution by graphic or tabular methods is always possible. The factors limiting the precision of such solutions were discussed by Romberg (1952). Computational procedures have been given by many authors; Hagedoorn (1954; 1959) discussed the problems from the standpoint of wave fronts rather than ray paths, an advance beyond the older methods.

The second degree of complexity in the interpretation of seismic data is to take into account changes in the velocity–depth relation from one shot point to the next. This is usually done by assuming that at a given shot point the velocity varies only with the depth, but that at neighboring points the velocity–depth relation is different.

The third degree of complexity is to assume that the velocity varies both vertically and horizontally at each shot point. This is of course the true state of affairs in the earth, especially if the formations dip. The iso-velocity surfaces will be neither horizontal, as in the simpler cases, nor will they be parallel to the bedding planes, since the velocity varies with both depth and lithology. The problem of determining the attitude of a reflecting surface from seismic data when the velocity varies horizontally has a solution (if the configuration is not too complicated) but it is usually not attempted in the field. To the knowledge of this reviewer no general study of seismic interpretation with horizontal velocity changes has been made.

Special types. Much attention has been paid to special types of interpretation. For instance, the near-surface layer has always been recognized as requiring special treatment because of its very low velocity (approximately 2000 feet/second as opposed to, say, 6000 feet/second at depths of 100 feet or so). Thralls and Mossman (1952) and Krey (1954) have dealt with this problem, which is fundamentally one of applying a correction. Diffractions, recognized long ago by Rieber (1937), are discussed by Krey (1952) and by Le Doux (1957). O'Brien (1957) and Bortfeld and Hürtgen (1960) deal with reflected refractions. Richards (1959; 1960) discusses broadside refraction shooting and wide-angle reflections in Canada. These special types are cited merely as examples of the way interpretive procedure is modified for special geologic problems. Many other examples might of course be mentioned, and descriptions of new ones appear regularly in the literature. They are valuable to the worker in exploration in that they broaden the scope of the problems he can solve.

Special applications. It has been pointed out that the great bulk of exploration seismology is carried on in the search for oil. Seismology in

oil exploration has become a routine high-speed operation that can make no attempt to get all the information about a prospect that could be extracted by the seismic method. The industry has held generally that the necessary effort is too costly to consider. The geophysicist who wishes to study an interpretation problem in more detail than is customary in ordinary oil prospecting must turn to some of the special applications. These are characterized by closer attention to the individual case, by devising field set ups especially tailored to the problem, by reshooting to check hypothetical interpretation, and by the invention of new computing techniques.

In oil exploration the industry has conceded that one place where special attention and specially devised techniques are important is in delineating of salt domes. It is important to understand as accurately as possible the shape of salt domes and the attitude of the beds they have deformed in the process of piercing them. The essence of the method is to use bore holes to get either the shot or the seismometer a significant distance below the surface. The elements of the method were described by Gardner (1949). Holste (1959) published a description of the mapping of salt-dome flanks and other interfaces in Europe. Musgrave and others (1960) gave a detailed description of the interpretation of a dome, including computing devices and model of the dome as interpreted.

Another problem requiring special attention is interpreting the results of shallow refractions and reflections in the search for ores and ground water as well as for corrections to deep exploration. Domzalski (1956) outlined the general problem of shallow refractions and investigated the limits of possible accuracy in the results. Pakiser and Warrick (1956) described the technique and results of shooting for shallow reflections at depths of the order of 100 feet, and Pakiser and Black (1957) reported on shallow refractions in uranium prospecting. Schmidt (1959) and Reichenbach and Schmidt (1959) reported on the results of seismic-reflection surveys in the siderite district in the Siegerland (West Germany) both on the surface and underground. Examples of the application to ground-water exploration are given by Berson and others (1959) and by Warrick and Winslow (1960).

GRAVITY IN EXPLORATION

General Features

Comparison with seismology. Gravity is next in importance to seismology as an exploration method. In practice nearly all exploration prospects are covered with a gravity grid before they are explored with the seismograph, although the station density of gravity stations varies widely according to the policies of different oil companies. Even with a fairly high station density, however, the cost of gravity per unit area covered is much less than that of seismology. The result is that, even though most prospects are covered by both methods, the manpower and money expended in gravity is less than one-tenth of that used in seismology. Nevertheless, many exploration men are familiar with seismology, whereas few are experienced in gravity.

There are two important differences between the intellectual processes involved in seismic and gravity exploration. The first is in the place where judgment is needed. In seismograph work judgment and originality are needed in the actual field work. The hole depth at which the shot is fired, the array geometry, and the filtering all need to be correctly determined on the spot in order to obtain good records. Once good records are in the hands of the interpreter, his duty is more or less prescribed, and he has little difficulty in reaching a reasonably unambiguous picture of the local structure. In gravity prospecting, on the other hand, field work does not require decisions; it is straight routine for which full instructions can be written. It is the interpretation which cannot be done by rules; each problem has to be solved in a different way from its predecessor. The second difference in the two methods is an outgrowth of the first. A seismograph interpretation, correctly performed, is the unique solution of a straightforward problem. A gravity interpretation, on the other hand, is never a unique solution of a problem. Many interpretations are always possible; the one that is chosen is a judicious guess instead of the result of a sequence of logical steps. This comes about because, although every configuration of masses gives rise to a unique gravitational field, any gravitational field observed at the surface can be caused by an infinite number of different mass distributions. Skeels (1947) illustrates this fact very clearly with examples and shows what a pitfall it can be to the superficial interpreter (Fig. 18).

Interpretation. Because it is fundamentally ambiguous, then, gravity can be interpreted only if the number of possible solutions of the problem can be reduced from infinity to a very few. This can be done if the local geology is well enough understood so the geophysicist can

decide what sort of structures he is looking for and develop a feeling for the kind of gravity anomalies they are likely to give. He is required to know not only the structural geology of his prospect but also the relative densities of the formations that are present and the methods for computing the gravitational attractions of

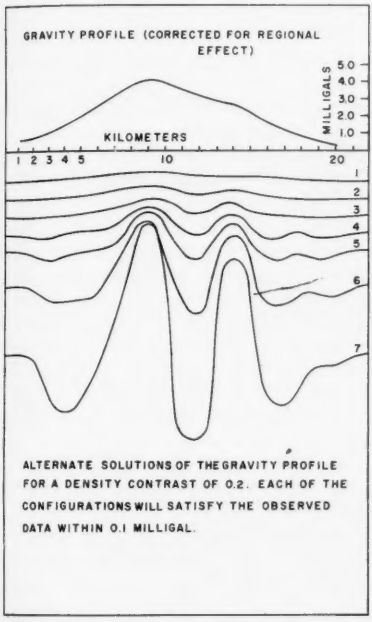

Figure 18. Gravity profile and basement configurations at various depths, each of which could give rise to it (After Skeels, 1947, Fig. 1)

the anomalous rock masses that are likely to be present. He must be aware of the fact that anomalies can be caused by lateral density changes that are not caused by structure at all. In practice the problems always have to be worked backwards. An anomaly of a given shape is isolated or defined. A hypothesis is evolved to explain what kind of structure may cause it. The attraction of such a structure is computed to see if its gravity anomaly resembles the observed one. If not, the hypothesis is modified until the computed anomaly fits the observed one within the limits of observational error. The interpretation thus consists of constructing the most likely and geologically reasonable hypothetical structure that will fit the observed data. To attempt to reduce this process to a set of instructions that can be followed without understanding what is really being done is to court failure. Much geologic information lies hidden in gravity maps today because the effort to interpret them was abandoned when a prescribed routine proved ineffective.

Field routine and data reduction. Unlike the interpretation, collecting gravity data in the field is routine and can be carried out by following prescribed rules. The value of gravity at a given station is found by comparing the reading of a gravity meter at the station with the meter reading at another station where the correct theoretical or Bouguer gravity is already known, and correcting the observed value for latitude, elevation, and the surrounding topography. In sum, a set of gravity data consists of the correct values of gravity, within known limits of error, for a set of locations. It is customary to plot these values on a map and to draw contours of equal gravity. Such a map is called a map of the Bouguer gravity anomaly. If its quality is known, it can then be interpreted with no further reference to the circumstances in which the data were obtained.

Defining the Anomaly

Regional anomalies. A spheroidal earth in which the density changed radially but not tangentially would have a gravity field on its surface which when mapped and corrected for latitude would be featureless. The features on an actual gravity map are caused by the major and minor irregularities in the earth's crust, such as changes in crustal thickness, mountains and mountain roots, and differences in composition in the deep and shallow layers of the crust. These irregularities cause all kinds of large and small ups and downs in the surface-gravity map and are thus of interest to the geodesist and the geologist because they are evidence of earth shape and continental and oceanic structure (Fig. 19). The exploration geophysicist, on the other hand, is concerned with ups and downs that range (relatively) from very small to extremely small. On the gravity map they look like minor hummocks and pockets on the sides of steep mountains, and they are not easy to see unless the map is processed in some way to make them visible.

Regional and residual maps. The way to extract the small gravity anomalies that are wanted for mineral exploration from the observed gravity is to construct a map of the so-called regional gravity and subtract the

values on it from the observed gravity. A regional gravity map is one that has been made smooth until no anomalies with horizontal dimensions smaller than an arbitrary minimum are present. A map that represents the difference between the observed and the regional gravity is called a "residual" map. It contains the small anomalies in a form in which they can be seen clearly and their size and shape determined.

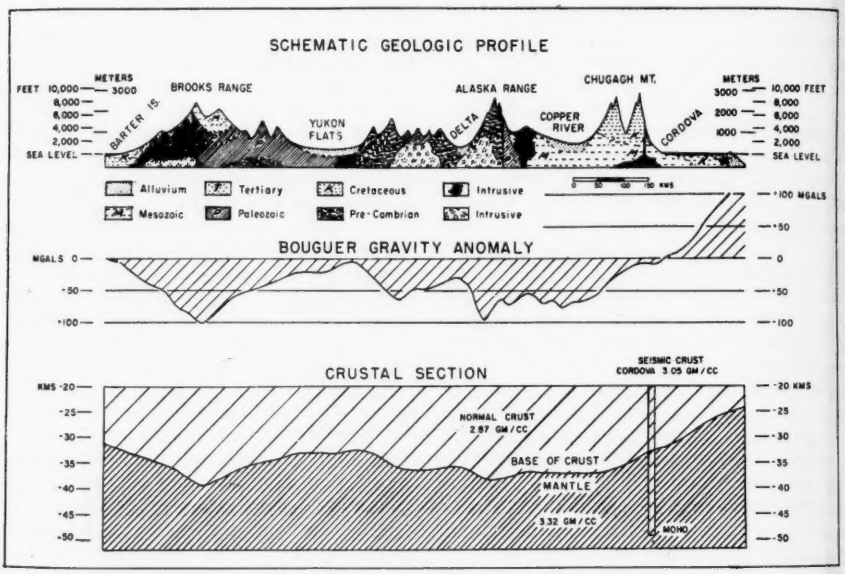

Figure 19. Relation of Bouguer gravity anomalies to geology, topography, and crustal thickness across Alaska (After Woollard, 1960, Fig. 3)

The only criterion used to separate "residual" anomalies from "regional" gravity is that of size—not magnitude, but horizontal extent of the smallest dimension of the anomaly. Thus several residual maps can be made from one set of gravity data, according to the anomaly width that is chosen as critical. The width of an anomaly (in simple cases) is proportional to the maximum possible depth of the mass that causes it. Thus a map of residual anomalies of a certain maximum width excludes the anomalies due to masses below a certain depth.

Residual anomalies. The preceding paragraph gave the theoretical or idealized way of finding residual anomalies. In practice, the pictures of residual anomalies are drawn in a number of ways, roughly divisible into graphical and mechanical. The primitive way is graphical. It consists of drawing profiles or cross sections in different directions across the map of observed gravity. If enough of these profiles are drawn, and their positions, orientations, and scale carefully chosen, they will contain the information that is in the original map. The anomalies are found by "smoothing" the profiles graphically. Anomalies of any desired width can thus be isolated, and "residual profiles" of the individual anomalies can then be constructed for detailed study. Drawing profiles and smoothing them is the best way for an interpreter to get acquainted with a gravity prospect, and the fact that it yields residual profiles of the anomalies directly is an advantage.

On the other hand, all graphical methods are laborious, and all processes of "smoothing" are necessarily somewhat subjective, which is a disadvantage when the skilled interpreter seeks to delegate his duties to people who have less experience. It would be advantageous to have an objective and routine method of con-

structing pictures of residual anomalies. Much effort has accordingly been put into devising mechanical or automatic ways of making residual maps. Baranov (1954) described an analytical method of finding the background or regional gravity by requiring that the square of the difference between the observed and the background value be a minimum. Grant (1957) and Krumbein (1959) presented more general solutions of the problem especially adapted to computer techniques. Fajklewicz (1959) gave a method using cracovian matrices for finding the regional gravity when computers are not available. Dean (1958) extracts the anomalies by a process of linear filtering analogous to the filtering in communication theory.

The relative merits of graphical versus mechanical residual maps are discussed by van Weelden (1953) in an excellent essay on the whole subject of gravity interpretation. He points out that graphical methods sometimes lead to insufficiently careful treatment because they are laborious, and that if they are not conscientiously applied some of the information contained in the data may be neglected. He shows, on the other hand, that mechanical methods are insufficiently flexible, because, for instance, in different parts of a basin the anomalies sought will be of different widths so that any one automatic method will not be sufficient to point up all the information. He concludes, in short, that both methods have their faults and to yield the best results must be used with full understanding of the physics and geology of the problem.

Second derivatives. One of the best mechanical methods of constructing a residual map is the second-derivative method. The essence of the second-derivative method is to find places where the curvature of the gravity map is sharp rather than where the anomalies are large or small. A *sharp* or *narrow* anomaly in the gravity is shown as a prominent anomaly in the curvature (ringed by a less prominent anomaly of the opposite sign); a *broad* gravity anomaly is minimized even though it may be very large. The purpose of the second derivative is thus to separate, by more or less automatic means, the narrow anomalies from the broad anomalies, emphasizing them in proportion to their *sharpness*, and minimizing them in proportion to their breadth. Elkins (1951), author of a generally quoted paper on the subject, said (p. 29): "The second derivative method of interpreting gravity data, although its use is justifiable only on data of high accuracy, offers a simple routine method of locating some types of geologic anomalies. . . ." Van Weelden (1953) comments that the limitations thus placed on the method are important and says: "With these limitations clearly in mind, the method may indeed in certain instances be helpful as a visual aid to spot smaller anomalies superimposed on larger effects." He also points out that Elkins' statement implied that the method was to be used for *locating* anomalies and not for *interpreting* them.

The importance of the second-derivative method of defining gravity anomalies is attested by its widespread adoption in the exploration industry and by the general acceptance among oil companies of the idea that studying a second-derivative map is the normal way to interpret gravity. Much has been published on the subject, beginning with Henderson and Zietz (1949a) who described second vertical derivatives for magnetics with a treatment equally applicable to gravity. A recent paper by Roland Henderson (1960) presents "a comprehensive system of automatic computation in magnetic and gravity interpretation" which will permit the computing by modern electronic digital computing equipment of any function of the gravitational field. Rosenbach (1954) studied the advantages of the second-derivative method in locating anomalies that are difficult to find on the map of observed gravity. Goguel and Lemoine (1957) point out that derivative maps are calculated from points, and that the ways of approximating second derivatives from points may give rise to errors in the interpretation. They say: ". . . if the result is interpreted as if it really were a second derivative, erroneous values for the depth and mass are obtained." It should be noted that the last paper makes it clear that second derivatives are regarded as an aid to interpretation as well as to the location of anomalies. This goes beyond the limitations quoted above from Elkins and is representative of a widely current view of derivative maps.

A paper published by Nettleton (1954) is probably still the most serious comparative study of derivative methods in the literature. In reviewing the subject Nettleton says: "numerous schemes have been used for treating regional effects and a certain aura of mystery and conflicting claims has come to surround these treatments." He compares graphical and mechanical methods at considerable length, naming their practitioners "smoothers" and "gridders" respectively, and giving a compre-

hensive review of published examples illustrating his comments. He points out that there is really no mystery in gridding methods, since they all consist of a comparison between gravity at a point and weighted averages from one or more rings of values surrounding the point. He shows that they distort anomalies to a certain extent, thus making it difficult to carry out proper computations for mass and depth. This aspect of interpretation by derivatives is further discussed by Romberg (1958) who points out that a second-derivative map is an incomplete representation of the data in that the magnitude, shape, and horizontal dimensions of anomalies cannot be recovered from it. It is therefore impossible for the interpreter studying it to establish quantitatively the physical connection between the geologic structure and its gravity. The obvious rule for a good interpretation is thus to locate all anomalies with the aid of a derivative map and then find out their true magnitude and shape from the map of the observed gravity. The observed-gravity map contains all the information that any derived map can show. If the derived map indicates an anomaly that cannot be discovered on the original, then the anomaly is a product of the process and is not really present.

Downward continuation. Attempts have been made to solve the problem of the gravitational field directly by making assumptions restrictive enough to make the direct solution possible. One way is to compute what the gravitational field would be at some level lower than the surface if it were not affected by odd masses between the surface and the chosen level. This is called the method of downward continuation and is often helpful in unscrambling the observed anomalies, although Hughes (1942) pointed out that the data had to be quite accurate for this purpose. Many methods of continuing the gravity field downward have been described; a recent example is the general treatment by Roland Henderson (1960). The physics of downward continuation can be explained in this way; if the gravity on the surface is all due to a mass distribution at a certain level below it (Tsuboi, 1938), that mass distribution can be calculated from the gravity. Geologically, a mass distribution at a given level can be imagined as structure at an interface where there is a density contrast, as from a sand section to an underlying limestone section. As the interface is thought of as being deeper and deeper, the anomalies in the observed gravity are found to come from individual masses or structures that are larger and more concentrated. As the level of mass distribution is assumed to be deeper still, the masses become absurdly large and have to be compensated with mass deficiencies beside them. Skeels (1947) illustrates the effect (Fig. 18). When such a depth is reached, it is concluded that the geologic picture is absurd and that the masses causing the anomalies are not really so deep. In nature, of course, the structures causing the anomalies are not all at one level. The maximum depth of the structure that causes each anomaly, in ordinary conditions, can be readily computed from the residual profiles of the anomaly itself, without the need of a series of downward continuation maps except possibly one for resolving the anomalies. The depth of the masses causing the anomalies is of course intimately related to the local geology, especially the density contrasts. On this account the anomalies should be studied individually.

Computations in Gravity

Attractions. After a gravity anomaly has been defined and its magnitude and shape determined as well as may be with the available data, the next step is to postulate a hypothetical mass of suitable depth, shape, and density so that its gravitational anomaly resembles the observed anomaly. This process is purely mechanical; given a mass and its dimensions and density, its gravity can be computed. Effective interpretation, however, depends on the ability to achieve, *in a reasonable time*, a close enough approximation to the desired gravity so the difference is not significant. If the time required is excessive, the interpretation may in practice not be carried out to the limit of its usefulness. Nettleton (1942) showed examples of simple shapes whose gravitational attractions could be used as first approximations in an interpretation problem. In most cases the shape of the observed anomaly is not known well enough so that more accurate computations are useful.

Many papers have been published that deal with methods of calculating the gravity and magnetic effects of postulated bodies, especially by computer and analogue means; the latest ones are cited here. Bott (1960) offered a computer program for computing the gravity of sedimentary basins, and Talwani and Ewing (1960) one for computing the attraction of arbitrary three-dimensional bodies. Gerrard and others (1957) and Roy (1959) described

optical analogue methods for making the computations.

Densities. An indispensable element in the solution of attraction problems is the relative density of the anomalous mass and of the surrounding material. In general the densities of sedimentary rocks vary within such wide limits that it is not possible to know the relative density of two masses correctly within 50 per cent unless samples are available, which is seldom the case. Densities are measured directly and indirectly in a variety of ways, but still not enough density measurements are actually made to insure good use of the data.

A general study of ways to measure density is presented by Whetton and others (1957). The effect of compaction on density—an important problem in countries where there is much relief for seismology as well as gravity—is discussed by Parasnis (1960). Domzalski (1955b) finds the density by making gravity readings in bore holes at different levels. Homilius and Lorch (1957) determine density by gamma-ray absorption. The use of gamma rays in getting density logs in bore holes is described by Pickell and Heacock (1960). There is much need for better data on the errors in these methods and for work in computing the over-all density contrasts of a section by correlating the geologic displacement with the observed anomaly.

Applications of Gravity

Many examples of the applications of gravity to problems in geophysical exploration have been published. It is not considered desirable to summarize these; the interested reader can find them in periodicals or in the Case Histories volumes (Nettleton, *Editor*, 1948; Lyons, *Editor*, 1956). Four different and unusual types of gravity prospects may be mentioned. Nettleton (1957) described the gravity over a mound on the continental shelf in the Gulf of Mexico; it is interpreted as a salt dome. Davis and others (1957) described the search for chromite deposits in Cuba. Domzalski (1955a) described a three-dimensional survey in a mine. DeBruyn (1955) showed a regional map of European and North African gravity.

MAGNETICS IN EXPLORATION

General Features

Magnetic prospecting is the oldest form of geophysical exploration. The use of compasses and dip needles in searching for iron ore was understood before the seismic, gravity, or electrical methods were thought of. Magnetic exploration is today more important in the search for ores than gravity or seismology, although it is definitely in third place in oil exploration. However, the new technique of aeromagnetic mapping has made it possible to get magnetic data faster and more cheaply than any other kind of geophysical data. This technique is useful in mapping broad sedimentary basins for oil exploration before other data are available.

Magnetics is of minor usefulness in small-scale oil exploration (in contrast to its importance in mining) because of the low magnetic susceptibility of most sediments. They are an order of magnitude below those of metamorphic rocks and two orders below those of basic igneous rocks; acidic igneous rocks are in between. It follows that in most petroliferous sediments the contrasts in susceptibility, which give magnetic anomalies in the same way as contrasts in density give gravity anomalies, are minute, and the resulting magnetic anomalies are hardly perceptible. One interesting exception is the Horse Thief shale in Montana. It has a large magnetite content—magnetic susceptibilities in sediments are almost entirely due to their magnetite content—and can be mapped with magnetic methods whenever it is close to the surface. Otherwise, in oil prospecting magnetic anomalies are usually assumed to be due to intrusions or to structure in the basement. In oil exploration, the basement means the boundary between sediments that may contain oil and metamorphosed or igneous rocks whose crystalline character precludes their being oil reservoirs.

The fact that sedimentary structures in oil provinces usually give inconspicuous magnetic anomalies is put to use in a reverse way. Gravity and magnetics are properly used together in oil exploration. If there is a gravity anomaly and no magnetic anomaly, it is deduced that a sedimentary structure is present. If a magnetic anomaly coincides with an observed gravity anomaly, it is deduced that there is an intrusion or an anomalous feature in the basement which may (other things being equal) cause both anomalies. Garland (1951) following Poisson (1826) pointed out that there is a relation between the gravitational and the magnetic potential of a body, which could be used in making certain deduc-

tions about it from the combination of gravity and magnetic data.

Interpretation

Comparison with gravity. It has been stated that the problem of what configuration of masses causes an observed gravitational or magnetic field is not soluble in the general case, because a given field on the surface can be due to an infinite number of possible configurations. If the configuration or distribution of mass occurs all at one level, however, the problem does have a solution. In the case of a gravity prospect it is usually unrealistic to assume that all the mass distribution is at one level, because density contrasts occur all up and down the sedimentary column. For some problems in magnetics the assumption can be fairly safely made, since susceptibility contrasts in the sedimentary column are weak, and most magnetic anomalies are due to susceptibility contrasts in the basement. This permits a direct solution to be found for the shape of the basement. The depths of the sources of individual closed anomalies can also be deduced from their horizontal dimensions, and if there are enough such anomalies the basement can be mapped, at least roughly.

In spite of this fact, the theory of interpreting magnetic data is closely allied to the theory of interpreting gravity data. Many papers on the subject (Nettleton, 1942; Henderson, 1960) deal with both branches of it. There are three main differences, aside from the practical ones just mentioned: (1) the direction of the earth's field is not always downward as in gravity, but varies in azimuth and dip; (2) the polarization of magnetic bodies is often not parallel to the earth's magnetic field because of remanent magnetism, which causes an additional complication in computing their magnetic fields; (3) magnetic bodies have positive and negative poles so that the magnetic field of a body is the field of its positive pole plus the opposite-pointing field of its negative pole. In gravity this would be like the field of a positive mass plus the field of a mass deficiency or negative density contrast below it. The consequence of this third difference is that, when the dimensions of a magnetic body are small compared to its depth, its magnetic effect decreases more nearly as the cube of the depth than as the square. The inverse cube law does not hold for large-scale bodies whose effective poles are a distance apart comparable to the distances to the point where the field is measured. In practice this has the result that the two important uses of magnetics are for observing features that are either much smaller than the features observed in gravity, or much larger. The smaller features are of course relatively shallow ore bodies, and the larger features are rock bodies in the basement whose equivalent poles are far apart.

Computations. The classic paper on interpreting magnetic data was written by Peters (1949). He presented a downward-continuation method for solving the field-to-configuration problem, taking advantage of the fact that there are no magnetic poles between the surface and the basement, to compute basement relief directly. He also showed methods for estimating the depths of sources of anomalies and he gave formulas for derivatives and field continuations. With airborne magnetics in mind, Henderson and Zietz (1949b) discussed upward continuations and showed that, in agreement with theory, no advantage was gained by flying at several levels. Vacquier and others (1951) developed the method that is now most commonly used in mapping basement relief from airborne data.

The problem of computing the magnetic anomalies caused by bodies of various shapes was studied by Henderson and Zietz (1957), Affleck (1958), Hutchison (1958), and Price (1959). Mooney (1952) described a method of measuring the magnetic susceptibilities of rock specimens, and Schriever (1958) constructed a device for doing the same thing in a nonfluctuating weak field, the condition in which they occur in the earth, since susceptibility may vary in strong alternating fields. Green (1960) discussed the complications that are introduced into interpretation by the remanent magnetization—the magnetization put permanently into rocks when they were formed, as opposed to that induced by the present earth's field.

Examples of magnetics in exploration. An example of the most important application of magnetic methods to exploration is given by Henderson and Zietz (1958) in general geologic interpretation of the total magnetic intensity map of Indiana. It is an advantage in petroleum prospecting, when a sedimentary basin is to be examined, to get a quick and relatively cheap map of its shape. A similar example is given by Steenland and Brod (1960); the area is small, and the target is uranium ores, but the problems and their solution are analogous.

Agocs (1958) and Agocs and Isaacs (1956)

published long one-dimensional magnetic profiles that show the regional geology clearly. Miller and Ewing (1956) show a regional seaborne magnetometer survey in the Gulf of Mexico and part of the Atlantic Ocean. Mining applications are illustrated by Zietz and Henderson (1955) with an aeromagnetic map of the Sudbury district in Ontario, which served as a test of the interpretation method and as a correlation with geology. A novel application is presented by Levanto (1959) who describes the design and application of a three-component magnetometer and its application in an iron mine in Finland.

ELECTRICAL METHODS IN EXPLORATION

General Features

Divisions. Electrical methods of exploration can be divided either by principle or by application. The principles are the observation of natural fields (self-potential and telluric), applied direct-current fields (resistivity and equipotential), and applied alternating fields (induced polarization and induction). The division of electric methods by application seems more natural.

The most important application of electrical methods in exploration is electric logging, which has become an indispensable tool in the production of petroleum. Next in importance is their application to mining, engineering, and the search for ground water, where they have been widely and successfully used. The least important use of electrical methods is in oil exploration on the surface.

Electric logging. Electric logging is closely related, physically and geologically, to surface measurements of the resistivity and self potential. As it happens, however, it is practiced by a different group of people and for a different purpose, as it is a branch of petroleum engineering. The most important purpose of an electric log of a bore hole is to detect the presence of oil; since deep wells are drilled with thick heavy mud, the oil tends to be pushed away from the hole and cannot be found by primitive methods of testing the well. A natural extension of this first purpose is to grade the sands for productivity and to find their boundaries and the boundary between oil and water for oil-well engineering purposes. Another application is that in suitable conditions electric logs greatly facilitate lithologic matching between neighboring wells.

Physically, the electric log consists of three or more electrodes disposed on a long weight (called a sonde) which is lowered into a bore hole. A voltage is applied between the surface and one of the electrodes, or between two of the electrodes, and a continuous recording is made of the potential differences between the other electrodes and a different spot on the surface, as the sonde goes down the hole. The details are given briefly by Nettleton (1940) and at greater length by Jakosky (1960). The result is a set of curves or graphs of the resistivity, plus a curve of the self potential. The object is to compute from these curves the resistivity of the interstitial water, the porosity of the formulation, and the resistivity of the rock itself, along the length of the bore hole. Wyllie (1960) describes the theory and practice of converting the observations to the desired quantities and explains the interpretation of the results and their geologic meaning. The relationships used in the process are approximate and subject to changes in conditions and to disturbing effects, so that a complete interpretation requires comparisons with nuclear and acoustic logs and much experience in allowing for disturbing factors. Chombart (1960) shows that the same is true to an even higher degree in interpreting the logs of carbonate reservoirs, because of the lack of regularity in the pore distribution. Doll and others (1960) attack the frontier of the logging problem and propose the use of new equipment such as induction-logging devices now in the process of development. The last three references are part of a symposium on logging techniques (Archie, *Special Editor*, 1960) published in the August 1960 issue of GEOPHYSICS. The symposium also includes an article on density logging (Pickell and Heacock, 1960) and one on radioactivity logging (Mardock, 1960).

An interesting innovation in well logging is described by Colombo and others (1959). The authors found a number of "organic oxidants" that react with petroleum, sulfur, mineral sulfides, and lignite. After a bore hole is logged in the ordinary way, the oxidants are added to the mud, and the hole is logged again. The self-potential logs are compared, and if differences between them are observed the presence of the desired minerals is postulated. Field tests have given results indicating the presence but not the quantity of the desired minerals.

Mining Applications of Electric Methods

Comparison with seismology and gravity.

Electrical methods are fundamentally better suited to mining than are seismology or gravity because of their scale. (To "mining" should be added other small-scale applications in engineering and ground-water search.) The seismic pulse with a frequency of 30 cycles per second has a wave length in typical country rock of 300 feet or so; searching for ordinary ore bodies with energy of this description is like looking for molecules with an optical microscope. The same kind of objections hold for gravity; ore bodies in general are too diffuse or not massive enough to be discovered with gravity. (Exceptions to both these rules have been mentioned.) Electrical methods have no such restrictions as to scale. In addition, ore bodies generally contrast well in resistivity or other electrical properties with the rock that surrounds them.

An excellent collection of examples of geophysics used in mining was published by the European Association of Exploration Geophysicists (1958). It includes 21 studies of geophysical surveys in mining, hydrology, and engineering.

Self-potential. The simplest method of prospecting with electric methods is the self-potential method. It is simple because it does not involve putting any current or voltage in the ground; it is only required to put electrodes into the ground and search for naturally occurring anomalous potentials between them. These potentials are caused by spontaneous electrochemical action in the neighborhood of an ore body. According to Sato and Mooney (1960) the potential is due to an electric current caused by the (p. 226) "reduction of oxidizing agents near the surface and oxidation of reducing agents at depth. The ore does not participate in either reaction, but serves as a conductor. . . ." The paper is a review of the whole subject of self-potential and contains generalizations about self-potential based on the study of a number of recorded cases. Self-potential anomalies are of the order of less than 1 volt. Strong anomalies have been reported for pyrite, pyrrhotite, chalcopyrite, chalcocite, covellite, graphite, and anthracite.

Resistivity and equipotential. Resistivity and equipotential methods are treated together here because they consist, in principle, of putting an electric current in the ground with two electrodes and measuring the potential differences in the neighborhood with two other electrodes. The difference lies in the fact that in resistivity measurements the electrode configuration is fixed, and in the equipotential method the electrodes are moved around and used as probes. The equipotential method measures the lines along which the current flows. If there is an anomaly in part of the field, then the lines, as mapped on the ground, will not be symmetrical. A description of the equipotential method is given by Dobrin (1960).

In the resistivity method a current is put into the ground with two electrodes, and the resultant voltage is measured between two electrodes that are in line with and between them. The results are a guide to (although not an exact measurement of) the resistivity of a layer in the earth whose depth is proportional to the electrode spacing.

The theory of interpretation of earth-resistivity data is extensively developed. The case of two, three, and four horizontal layers was dealt with by Mooney and Wetzel (1956), and other authors have taken up the more complicated cases. Roman (1959) applied image analysis to the multi-layer problem, and Alfano (1959) developed a general theory for structures bounded by vertical and horizontal planes. An application is reported by Cook and Van Nostrand (1954), but in general more theory than examples appears in print.

Induced polarization. In induced polarization two electrodes are put into the ground, and a voltage is applied across them. After a current has been set up, the voltage is cut off. The current will continue to flow for a short while after the cut-off, apparently because an electro-chemical reaction has been set up and can be observed with two nearby electrodes. What happens is roughly analogous to the charging and discharging of a storage battery; the essential elements are a metallic ore body and an electrolyte. The reaction differs for different surroundings, so that, if an exploration target such as pyrrhotite is present, the current lasts longer than in ordinary earth, and its decay curve will have a different appearance from the usual one. Anomalies in the decay curves can be correlated with anomalies in the soil.

An account of the theory and practice of induced polarization was published by Bleil (1953). Vacquier and others (1957) showed that the effect existed where clay minerals were present and described a system for using it to prospect for ground water; the authors made limited claims for the method and recommended its use as a supplement for resistivity methods. Frische and von Buttlar (1957) presented a mathematical solution for the depth of an aquifer in the case of a nonpolarizable upper layer and an underlying polarizable layer. Šume (1959) gave some good examples

of induced polarization in mining. The causes of the phenomenon were studied by Marshall and Madden (1959), and a mathematical formulation was proposed by Siegel (1959). McEuen and others (1959) measured the electrical properties of synthetic specimens of targets of induced polarization prospecting.

Electromagnetic induction. Electromagnetic induction uses the principle of the transmission of radio waves. If a loop of wire is set up and an alternating current is made to flow through it, an electromagnetic field exists around it. If an electrical conductor (such as a radio antenna) is in the neighborhood the alternating field will induce currents to flow in the conductor, and these currents will set up an electromagnetic field of their own, which can be observed if it is strong enough. The anomalous currents have the same frequency as the energizing current but can be detected because they are out of phase with it. The method has long been used in exploration, especially mining, and is described in books (Dobrin, 1960, p. 366–369). Törnquist (1958) gives an example of the use of electromagnetic prospecting in mining work, describing a two-airplane method of looking for conducting ore bodies. A modern theory of the effect of an alternating field on a spherical ore body is given by Wait (1960), who showed that both electric and magnetic modes of the response should be considered. Ward (1959a) showed that, by varying the frequency and combining the results with gravity, the characteristics of a spherical ore body could be found uniquely. The theory and practice of the electromagnetic method in looking for ores are, as in gravity, widely separated. In order to make a theory workable the cases must be reduced to great simplicity— not always a good description of nature.

Electrical Methods in Oil Exploration

History. From the beginning of geophysical exploration a persistent effort in research has been applied to the problem of using electrical methods in oil prospecting. Electrical energy is so easy to handle, both in practice and theory, that its adaptation to oil exploration has been attempted repeatedly. It seems to be unsuitable because it is attenuated so rapidly in the earth that its use in exploring deep structures is not practical.

The use of electromagnetic energy for reflections was considered by Yost (1952) who developed a general theory for it. In the same year Pritchett (1952) measured the attenuation of electromagnetic waves in the earth and found it so high as to preclude their general use as exploration tools for oil. McGehee (1954) made an analogous experiment in the Carlsbad Caverns and found that the attenuation was in good agreement with theory—that is, it was too high to allow much hope in oil exploration. Orsinger and van Nostrand (1954) made an experimental determination of the depth to the Austin chalk–Eagle Ford shale contact in central Texas. The resistivity of the former is 20–40 ohm-meters and of the latter 5 ohm-meters, so there is a good contrast. The results, for depths down to 400 feet, were good to within 5 per cent. The theory of what happens at a horizontal resistivity contact was discussed by Slichter and Knopoff (1959).

Telluric currents. An exception to the rule that electrical methods have not been successfully applied to oil exploration is the use of telluric currents. Telluric currents are currents flowing naturally through the earth on a broad scale. They can be detected with pairs of electrodes properly arranged to cancel the effects of self-potential between the electrodes and the earth. Large-scale anomalies in the conductivity of earth materials, such as salt domes, cause measurable anomalies in the telluric currents. Boissonas and Leonardon (1948) outlined the Haynesville dome in Texas with telluric currents. Tuman (1951) investigated the limits of applicability and resolution of the telluric method and stated that depth estimates based on it are not reliable. Mainguy and Grépin (1953) reported the successful application of the method in four different areas in France. They commented that the method had not had widespread application in the U.S.A. because the search there was for small structures at great depth, to which the telluric method was not suited.

Cagniard (1953) published a new theory for the use of telluric currents which involved measuring the magnetic as well as the electric field and using Maxwell's equations to derive his conclusions. He suggested that the method could be used to measure the depths of large sedimentary basins. A later article (Cagniard, 1956) on telluric electricity is a basic treatment of electricity in the earth.

Vinogradov (1959) observed vertical telluric currents in Lake Baikal in Siberia.

RADIOACTIVITY IN EXPLORATION

An important branch of exploration geophysics, which has grown up since World War II, is the search for radioactive minerals. This search has two goals: one is the discovery of

fissionable materials for nuclear power in weapons and generators, and the other is the discovery of rare minerals and metals useful in alloys which happen to be associated with uranium, such as beryllium, columbium, tantalum, and the rare-earth metals. The search for radioactive minerals is unlike the search for oil, aquifers, sulfide ores, and most other targets of exploration, because it is direct. The radioactive minerals give off radiation, which is picked up by a Geiger-Müller counter or a scintillation counter. These counters are carried (or flown) across country according to a scheme calculated to cover it adequately; if an increase in radiation is noted, the search is intensified locally until its source is found. The relation of radioactive minerals to each other and to the rocks in which they occur is an important branch of geology, and the subject of radioactive decay and the construction of radiation counters are important branches of physics. The two, however, have much more of a separate existence than in other kinds of exploration, so that the frontiers of the problem lie separately in geology and in physics, and not in new combinations of the two. An exception to this generalization is pointed out by Broding and Rummerfield (1955) who describe the simultaneous use of gamma-ray and resistivity logging in the search for uranium.

An exposition of the fundamentals of radioactivity in exploration is given by Dobrin (1960, p. 374–397) who lists relevant publications in geology and physics. Recent writing on the subject includes a paper on the gamma-ray spectrometer by Mero (1960). The gamma-ray spectrometer, unlike ordinary radiation counters, does not merely add the total radiation but separates out the different energy levels. It can accordingly be used as an assaying or analyzing device for radioactive ores, especially in bore holes. Airborne radioactivity surveys and their applications were discussed by Moxham (1958; 1960).

INSTRUMENTS FOR EXPLORATION

Invention and Design

The invention and design of instruments for geophysical exploration have proved to be more the province of the physicist and the engineer than of the geologist. With some exceptions the technology of constructing geophysical instruments is at present adequate to its task, and the frontiers of the art consist in making the instruments more convenient and more reliable for field use, rather than making them so they will take measurements that are finer or of new types. Whenever a new requirement is set up, as for an airborne scintillation counter or a high-frequency seismic system, the requirement is promptly filled by the industry; the problem is not to build the instrument but to interpret the data. One exception to this rule is the task of building gravity meters that will operate in a moving vehicle such as a ship or an airplane. Even in this example, however, development of the meter has outstripped the ability of the geophysicist to determine the velocity of the vehicle. Another possible exception is a machine to determine elevations, which has not yet been brought to a point where it replaces the surveyor. Yet this is not wholly an exception, because surveying can be done with full adequacy without any machine at all, so an elevation meter would be merely a convenience. Gravity meters on land can measure gravity to a far greater accuracy than is useful in the field, because of the irregularity of the background or "noise level"; magnetometers have more potential accuracy than is necessary. Seismographs are now built that are sensitive enough so that they measure ground motion anywhere, so that higher sensitivity would only amplify the ground motion further. Because the problem of instrumentation is not a geophysical problem in the same sense as the interpretation of seismic or gravity data, it will not be discussed at length. Born (1960) gives a treatment of instrumentation suitable as a guide for further study.

Seismograph Systems

Seismometers. Seismometers (geophones, detectors) used in the portable field systems required for exploration are nearly always simple oscillating devices consisting of a weight supported by a leaf spring, with electromagnetic or oil damping. The damping is usually less than critical. The natural frequency is usually 15 cycles or higher, in order to discriminate against surface waves, although instruments used for refractions have frequencies as low as 2 cycles. Their output is a voltage caused by the relative motion of a magnet and a coil or by the changes in magnetic flux through a coil due to changes in the air gaps of a magnetic circuit. The voltage is thus proportional to the velocity of the earth's motion; usually no attempt is made to deduce the actual ground displacement

from the seismometer signal. A basic discussion of geophone design was published by Dennison (1953), and a description of seismometers for scientific use is given by Coulomb (1956). Dennison (1960) examines the theory of the response of seismometers to incident energy.

The main problem with seismometers was formerly to get a signal strong enough so that amplifier noise did not interfere with its recording. Now that amplifiers are very much less noisy than they were, seismometers can be made much less sensitive and therefore lighter, so that they now weigh a pound or less. Linear reproduction is not considered important.

Much seismic exploration has been done in bays, gulfs, and even in oceans where the bottom is shallow enough for drilling. For this kind of work it has been found appropriate to use seismometers that respond to pressure rather than to velocity, as they must be submerged to pick up the reflections through the water. The sensing element of these pressure seismometers is not a moving system but a crystal (barium titanate, or, recently, lead-zirconate-titanate) whose electrical characteristics change with the pressure. The technical literature of geophysics does not contain descriptions of pressure-phone seismic systems. However, the art of using these so-called piezoelectric crystals is well known in underwater-sound engineering. A text on the subject was published by Mason (1950), and a recent article by Sims (1959) describes a calibration hydrophone that is a good example.

Beckmann and others (1959) described a seismological application of ordinary fathometric echo sounding; their apparatus showed reflections not only from the ocean bottom but from interfaces more than 1000 feet below it.

Amplifiers and filters. The design of amplifiers for use in seismic systems is mainly a problem in electrical engineering. The amplification of which an amplifier is capable is a function of the freedom from noise in its components, and modern technology has done much to eliminate what was once inevitable noise in amplifier circuits. If a specified gain, and freedom from noise, and freedom from harmonic distortion, is desired, it can be provided. The problem is to provide it cheaply and in a light package for field use. Hermont (1956) gives a list of the requirements for a satisfactory amplifier system.

Filters are a somewhat more controversial problem, since they inevitably distort the transient pulses that pass through them. The extent to which such distortion, for the sake of excluding noise made by wind or unwanted energy in the earth, is permissible is properly a problem in geophysical interpretation and is mentioned in the section on Filtering. Once the geophysicist has decided the degree of filtering that is optimum for his purpose, this filtering can be given to him by the engineer.

An important part of the amplifier-filter system is the automatic gain-control system for holding record-trace amplification to a more or less constant level, so that events of small amplitude are not overshadowed by stronger events. The standard method has been to control gain from the current average-trade intensity, and this causes distortion. Merlini (1960) describes a method for excluding distortion by programming the gain control, so that events have the proper amplitude relation to their immediate neighbors.

Automatic data processing. The most promising avenue of development in seismic instrumentation today is automatic data processing. Automatic data processing is definitely a problem in geophysics, because the information that the geophysicist or the geologist needs cannot be requested of the physicist or engineer in set form but must be worked out mutually between the two professions. The most sophisticated processing is no improvement unless it gives geological information (not merely purity of signal) not previously available; what is needed is to get a type of processing that will give new kinds of data. The subject has been mentioned, and some recent publications cited. Loper and Pittman (1954) described seismic recording on magnetic tape, a technique, or type of technique, essential to advanced data processing.

Gravity Meters

Airborne gravity measurements. The latest development in gravity meters is the successful use of a gravity meter in an airplane. Nettleton and others (1960) and Thompson and LaCoste (1960) report airborne gravity measurements that are apparently correct to $\pm$ 10 mgal, when compared with ground gravity. The limitation on accuracy was not in the gravity readings themselves but in the probable error in the velocity of the vehicle. Airborne gravity measurements are thus too coarse to be used for exploration, where anomalies may be much less than 10 mgal, but it will have widespread applications in geodesy.

Gravity measurements at sea. Gravity meas-

urements with meters on surface ships have been reported by LaCoste (1959), Harrison (1959), and Worzel (1959). Harrison's work indicated a spread of about 2 mgal in the results after a small systematic error had been removed. The method appears to have superseded the older pendulum method, as used in submarines, and, although it is apparently not as accurate as measuring gravity with meters on a submarine (Spies and Brown, 1958) because of the accelerations, it promises to be good enough for geodetic purposes.

Gravity meters in motion. Gilbert (1949) described a gravity meter designed to be independent of motion. Dolbear (1959) adapted this design, known as the Gilbert string gravity meter, to a bore-hole meter. Lozinskaya (1959) and Fedynsky (1959) dealt with the subject theoretically. No reports have appeared of an actual model of the string gravity meter that promises the accuracy necessary for exploration.

Magnetometers

Airborne. Instruments for measuring the intensity of the earth's magnetic field for prospecting purposes were, until World War II, devices for balancing the earth's magnetic field against gravity. They had therefore to be leveled and could not be used in a moving vehicle, since the accelerations of a moving vehicle and the acceleration of gravity are not distinguishable by the instrument. The problem of overcoming the effect of motion and providing a continuous moving magnetic record was solved with the aid of a flux-gate element developed by Victor Vacquier and described by Muffly (1946) and by Wyckoff (1948). The instrument was used by the Navy as a submarine detector in World War II, and shortly afterwards was adapted to geophysical exploration. It has since completely altered the role of magnetic data in geophysics by providing fast coverage over distant or inaccessible areas.

Proton precession magnetometer. The proton precession magnetometer is an instrument for observing the effect of variations in the magnetic field on elementary charged particles. It was described by Packard and Varian (1954) and by Waters and Phillips (1956). A recent discussion is given by Hurwitz and Nelson (1960). It is well adapted to airborne work and is in principle capable of much higher accuracy than the older instruments.

Bore-hole magnetometer. The bore-hole magnetometer described by Levanto (1959) has already been mentioned in its application to problems in mining.

AFMAG. A new application of the magnetic field to exploration is the so-called AFMAG or alternating fields magnetic method, mentioned by Slichter (1955b) and more fully described by Ward (1959a). The previously existing limiting factor of airborne electromagnetic prospecting was the signal-to-noise ratio. In order to penetrate deeply enough into the rocks with the electromagnetic method, the receiving and transmitting coils had to be so far apart that the noise obscured the signal. The alternating-field method uses the noise itself—described as "natural magnetic fields in the audio and sub-audio range" due to thunderstorms and to reactions of various kinds in the ionosphere. The tilt of the plane of polarization of these fields is affected by magnetic anomalies.

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Late Pleistocene Climate of Europe: A Review

Abstract: The Pleistocene periglacial environment of Europe is believed to have been marked by intensive frost action and wind action and by the spread of arctic and subarctic plants and animals south to the Alps. The frost action is recorded by ice-wedge casts, involutions, solifluction deposits, block fields, and related deposits, some of which may record permafrost. The occurrence of extensive sand dunes and loess deposits may be related to the distribution of glaciofluvial plains, strong winds, and the absence of forest cover. The occurrence of a tundra zone in the basal levels of Late-glacial pollen sequences in central Europe indicates a frigid climate, and molluscan and mammalian faunas provide similar information.

Poser and Büdel have separately attempted to represent the Pleistocene climatic-vegetational zones of Europe on maps. Poser located the southern limit of permafrost according to the distribution of ice-wedge casts and involutions, and the northern limit of forest from pollen-analytical evidence summarized by Firbas. These two map lines were then used to help delineate several vegetational-climatic provinces. An additional study on the Late-glacial sand dunes led to the presentation of a map of mean summer air pressure for Europe.

Büdel located the Pleistocene frost limit on the basis of its presumed relation to the 1000-m snow line, which is identifiable in the mountains of central Europe. His representation differs from Poser's chiefly for Hungary and France, where there is a dearth of pollen studies from which vegetational relations can be inferred. Büdel distinguishes several vegetational zones throughout Europe on the basis of loess distribution and also from pollen diagrams where available. Büdel believes that the ice sheet had only a minor effect on the climatic zonation of Europe during the Pleistocene.

Although exception is taken to many details of the criteria used by Poser and Büdel for vegetational zonation, the maps are considered useful as a basis for discussion of Pleistocene atmospheric circulation patterns.

The sequence of climatic changes for the last cold period is inferred largely from the stratigraphy of periglacial loesses and intercalated soils, and from their correlation with the Alpine glacial sequence. The controversy concerning the classification of the Würm period as single (following Penck) or as multiple (following Soergel) is reviewed; the validity of the Göttweig interstadial is questioned on the basis of recent field studies and radiocarbon dates in Austria and adjacent areas.

The reduction of mean annual temperature during the late Pleistocene, as inferred from the snow-line depression in the mountains, amounted to 5°–8°C under the general assumption of a 0.5°–0.7°C/100 m vertical temperature gradient (lapse rate) in the atmosphere. Calculations based on the occurrence of frost features and fossil plants in lowlands, however, suggest a temperature depression of 10°–12° C. The discrepancy may result from different lapse rates over mountains and over lowlands, according to Mortensen. Studies of modern lapse rates in arctic, subarctic, and subtropical regions are reviewed in order to evaluate this hypothesis.

Maps of mean atmospheric pressure and circulation patterns are commonly based on extrapolation from those modern patterns that are related to snowy winters, cool summers, and other conditions favorable to glaciation, on the assumption that the Pleistocene climate was marked primarily by a different frequency of certain air-mass movements and related storm tracks compared to today. Possible causes for circulation changes are still highly speculative, although much recent interest is centered on variations in solar radiation such as are exhibited in sunspots.

Zusammenfassung: Es wird allgemein angenommen, dass die periglaziale Zone Europas während des Pleistozäns starker Frost- und Windwirkung ausgesetzt war, infolge welcher arktische und subarktische Pflanzen sich damals in südlicher Richtung auf die Alpen zu verbreiteten. Die Frostwirkung zeigt sich in Eiskeilnetzen, Taschenböden, Fliesserden, Blockfeldern und ähnlichen Ablagerungen. Einige davon sind Anzeichen eines Dauerfrostbodens. Das Vorkommen ausgebreiteter Flugsand- und Lössablagerungen muss an der Verbreitung fluvioglazialer Flächen, starken Winden und dem Fehlen einer Waldbedeckung liegen. Die Pollenspektren in den untersten Teilen spätglazialer Ablagerungen in Mitteleuropa weisen auf ein kaltes Klima hin, ebenso wie die Mollusken und Säugetiere.

Unabhängig voneinander haben Poser und Büdel versucht, Karten der klimatisch bedingten Vegetationsgürtel Europas während des Pleistozäns

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zusammenzustellen. Die Südgrenze des Dauerfrostbodens ist von Poser auf Grund der Verbreitung von Eiskeilnetzen und Taschenböden gezogen worden, während die Nordgrenze des Waldes mit Hilfe der pollenanalytischen Daten nach Firbas festgestellt wurde. Diese beiden Linien wurden benutzt, um die verschiedenen Vegetations- und Klimaprovinzen abzugrenzen. Eine ergänzende Untersuchung spätglazialer Binnendünen hat zu der Zeichnung einer Karte der mittleren Luftdruckverteilung während des Sommers in Europa geführt.

Büdel hat die pleistozäne Frostgrenze auf Grund der angenommenen Beziehung zur 1000 m-Schneegrenze festgestellt, die in den Mittelgebirgen nachgewiesen werden kann. Seine Darstellung weicht von der Posers hauptsächlich für Ungarn und Frankreich ab, wo noch wenig pollenanalytische Untersuchungen zur Bestimmung der Vegetation durchgeführt worden sind. Auf Grund der Lössverbreitung sowie auf Grund von Pollendiagrammen, wo solche zur Verfügung standen, unterscheidet Büdel verschiedene Vegetationszonen in Europa. Büdel meint, dass das Inlandeis nur einen untergeordneten Einfluss auf die Klimazonen Europas während des Pleistozäns ausgeübt haben kann.

Obwohl man in manchen Einzelfragen Bedenken gegen die von Poser und Büdel für die Vegetationszonierung benutzten Normen haben kann, müssen diese Karten doch als sehr wertvoll für die Diskussion der Luftdruckverteilung und Windrichtung während des Pleistozäns beachtet werden.

Die Aufeinanderfolge der Klimaschwankungen während der letzten Eiszeit wird hauptsächlich aus der Stratigraphie der periglazialen Lössschichten und der eingeschalteten Böden einerseits und aus ihrer Verknüpfung mit den alpinen Vergletscherungen andererseits abgeleitet. Die Streitfrage über die Würmeiszeit als ein einmaliger (nach Penck) oder mehrmaliger (nach Soergel) Kältevorstoss wird diskutiert, während die Haltbarkeit des Göttweiger Interstadials auf Grund von rezenten Untersuchungen im Gelände und von Radiokarbon Datierungen in Österreich und in angrenzenden Gebieten in Frage gestellt wird.

Unter der üblichen Annahme eines Reduktionsfaktors von 0.5 bis 0.7° C. für 100 m lässt sich aus der Senkung der Schneegrenze im Gebirge eine Erniedrigung der mittleren Jahrestemperature von 5 bis 8° C. berechnen. Aus dem Vorkommen von Frosterscheinungen und fossilen Pflanzen im Flachland lässt sich dagegen ein Temperaturrückgang von 10 bis 12° C. berechnen. Nach Mortensen könnte diese Diskrepanz ungleichen Reduktionsfaktoren im Gebirge und im Flachland zugeschrieben werden. Um diese Hypothese auszuwerten, werden in diesem Aufsatz die Untersuchungen über den jetzigen Reduktionsfaktor in arktischen, subarktischen und subtropischen Gebieten besprochen.

Karten der mittleren Luftdruckverteilung und der Windrichtungen sind meistens auf einer Extrapolation derjenigen rezenten Luftdrucklagen basiert, die schneereiche Winter, kühle Sommer und andere für Vergletscherung günstige Bedingungen hervorrufen. Dabei wird angenommen, dass das pleistozäne Klima im wesentlichen durch eine von der gegenwärtig abweichenden Häufigkeit von Strömungen bestimmter Luftmassen und damit in Verbindung stehender Sturmbahnen charakterisiert war. Über die möglichen Ursachen der Änderungen des Strömungsbildes lässt sich noch wenig sagen, obwohl sich das heutige Interesse besonders auf die Schwankungen der Sonnenstrahlung, wie sie in Sonnenflecken zum Ausdruck kommt richtet.

Résumé: On considère en général que, pendant le Pleistocène, le milieu périglacial d'Europe était caractérisé par une action du gel intense, par l'action du vent, et par l'extension vers le sud, jusqu'aux Alpes, de plantes et d'animaux arctiques et subarctiques. L'action du gel est indiquée par des fentes en coin, par des involutions, par des dépôts de solifluxion, par des champs de blocs et autres dépôts analogues, dont certains peuvent indiquer un sol gelé en permanence. La présence de dunes sableuses importantes et de loess est peut-être liée à la répartition de plaines fluvioglaciaires, à des vents violents, et à l'absence de couvert forestier. La présence d'une zone de toundra dans les niveaux inférieurs des diagrammes polliniques de l'époque Tardiglaciaire de l'Europe Centrale indique un climat froid, et les faunes de mollusques et de mammifères fournissent des renseignements semblables.

Poser et Büdel ont indépendamment essayé de représenter sur des cartes les zones de climat et de végétation de l'Europe au Pleistocène. Poser place la limite sud du sol perpétuellement gelé d'après la répartition des fentes en coin et des involutions, et la limite nord de la forêt d'après les indications de l'analyse pollinique telles qu'elles sont résumées par Firbas. Ces deux lignes sont ensuite utilisées pour permettre de séparer plusieurs provinces de climat et de végétation. Une étude supplémentaire des dunes de sable du Tardiglaciaire permit de dresser une carte de la pression atmosphérique moyenne d'été pour l'Europe.

Büdel détermine la limite de gel du Pleistocène d'après ses rapports avec la limite des neiges permanentes de 1000 m, qui se reconnait bien dans les montagnes de l'Europe centrale. Sa représentation diffère de celle de Poser surtout en Hongrie et en France, où les études polliniques sont insuffisantes pour qu'on puisse en déduire des renseignements sur la végétation. Büdel distingue plusieurs zones de végétation dans toute l'Europe d'après la répartition du loess et aussi d'après des diagrammes polliniques lorsqu'ils existent. Büdel pense que la calotte glaciaire n'a eu qu'un effet

minime sur la zonation climatique de l'Europe au Pleistocène.

Bien qu'on puisse critiquer bien des détails des critères utilisés par Poser et Büdel pour la zonation de la végétation, leurs cartes se montrent utiles comme bases de discussion des systèmes de circulation atmosphérique du Pleistocène.

L'auteur déduit la succession des changements de climat pour la dernière période de froid et général d'après la stratigraphie des loess périglaciaires et des sols intercalés, et de leur corrélation avec la succession glaciaire des Alpes. Il résume la controverse qui entoure la classification de la période du Würm en une seule glaciation (suivant Penck) ou en glaciations multiples (d'après Soergel); il met en question la validité de l'inter-stadiaire du Göttweig en vue des études récentes sur le terrain et des dates au radiocarbone obtenues en Autriche et dans les régions avoisinantes.

L'abaissement de la température moyenne annuelle pendant le Pleistocène récent, déduite de l'abaissement de la limite des neiges dans les montagnes, est de 5 à 8°C, si l'on admet l'existence d'un gradient vertical ("lapse rate") de température de l'atmosphère de 0,5 à 0,7°C par 100 m d'altitude. Toutefois les calculs basés sur la présence de formes de cryoturbation et de plantes fossiles dans les plaines suggèrent un abaissement de la température de 10 à 12°C. La différence provient peut-être, d'après Mortensen d'une différence entre les taux de décroissement au-dessus des montagnes et au-dessus des plaines. L'auteur résume les études des taux actuels dans les régions arctiques, subarctiques et subtropicales pour évaluer cette hypothèse.

En général, on dresse les cartes de pression atmosphérique moyenne et des systèmes de circulation par extrapolation à partir de ceux des systèmes contemporains qui se trouvent liés à des hivers neigeux, des étés frais et autres conditions favorables à la glaciation, ceci étant basé sur la supposition que le climat Pleistocène était caracterisé surtout par une fréquence de certains mouvements des masses d'air et des trajectoires de tempêtes différente de ce qui se passe de nos jours. Les causes possibles du changement de circulation sont encore du domaine de la conjecture, quoiqu'on s'intéresse actuellement beaucoup aux variation du rayonnement solaire telles qu'elles s'observent dans les taches du soleil.

Resumen: El medio ambiente periglacial del Pleistoceno de Europa se cree que ha sido marcado por los efectos intensos de heladas y de vientos y por la dispersión de plantas y animales árticos y subárticos hacia el sur, hasta los Alpes. El efecto de las heladas está registrado por los moldes de cuñas de hielo, por involuciones, por depósitos de soliflucción, por campos de bloques y por depósitos relacionados, algunos de los cuales pueden registrar la helada permanente. La presencia de extensos médanos arenosos y depósitos de loess puede estar relacionada con la distribución de llanuras glaciofluviales, con fuertes vientos y con la falta de cubierta forestal. La apariencia de una zona de tundra en los niveles basales de los diagramas de polen hacia el final de la glaciación en Europa central indica un clima frígido. Las faunas de moluscos y de mamíferos proporcionan datos parecidos.

Poser y Büdel, por separado, han intentado representar en mapas las zonas climático-vegetacionales pleistocénicas de Europa. Poser localizó el límite austral de la helada permanente de acuerdo con la distribución de moldes de cuñas de hielo y de involuciones, así como el límite boreal de los bosques de acuerdo con las evidencias analíticas de polen resumidas por Firbas. Estos dos límites fueron aprovechados para delinear varias provincias climático-vegetacionales. Un estudio adicional de los médanos arenosos del final de la glaciación dió como resultado la presentación de un mapa de la presión atmosférica media estival de Europa.

Büdel localizó la línea de las heladas pleistocénicas a base de las relaciones supuestas con la línea de nieve de 1,000 m., que es identificable en las montañas de Europa central. Su representación difiere de la de Poser principalmente para Hungría y Francia, donde escasean los estudios de polen de los que se podría inferir las relaciones vegetacionales. Büdel distingue varias zonas vegetacionales en toda Europa a base de la distribución de loess y también de los diagramas de polen disponibles. Büdel cree que el casquete de hielo sólo tuvo un efecto menor en la zonificación climática de Europa durante el Pleistoceno.

Aunque se duda de muchos de los detalles de los criterios usados por Poser y Büdel para la zonificación de la vegetación, los mapas se consideran útiles como una base para la discusión de los regímenes de circulación atmosférica del Pleistoceno.

La sucesión de los cambios climáticos en el último período frío se infiere principalmente de la estratigrafía de los loess periglaciales y suelos intercalados, así como de su correlación con la sucesión glacial alpina. Se hace una reseña de la controversia acerca del período Würm, clasificándolo como uno solo (siguiendo Penck) o como múltiple (siguiendo Soergel); se pone en duda la validez del interestadial de Göttweig, basándose en estudios recientes de campo y en fechas radiométricas de carbón procedentes de Austria y de áreas adyacentes.

La disminución de la temperatura media anual durante el Pleistoceno tardío, inferida de la depresión de la línea de las nieves en las montañas, alcanzó de 5° a 8° C., bajo la suposición general de un gradiente vertical de temperatura en la atmósfera de 0.5° a 0.7° C. por 100 m. Los cálculos basados en la presencia de rasgos de heladas y de plantas fósiles en las tierra bajas, sin embargo, sugieren una disminución de la temperatura de

10° a 12° C. La discrepancia puede deberse a diferentes relaciones de tardanza que existían en las montañas y en las tierras bajas, de acuerdo con Mortensen. Con el objecto de evaluar esta hipótesis, se hace una reseña de los estudios de las relaciones modernas de tardanza en las regiones árticas, subárticas y subtropicales.

Mapas de presión atmosférica media y de regímenes de circulación atmosférica generalmente se basan en la extrapolación a partir de los regímenes modernos que están relacionados con los inviernos nevados, los veranos frescos y otras condiciones favorables a la glaciación, suponiendo que el clima pleistocénico fue marcado sobre todo por una frecuencia diferente de la actual de ciertos movimientos de masas de aire y de los cursos de las tormentas relacionadas. Las causas posibles de los cambios de circulación son aún altamente especulativas, aunque recientemente se ha enfocado mucho interés en las variaciones de radiación solar, como las exhibidas por las manchas solares.

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Резюме: Существует мнение, что перигляциальные условия во время плейстоцена были отмечены в Европе усиленным действием мороза и ветра, а также распространением арктических и субарктических растений и животных к югу от Альп. На действие мороза указывают погребенные слепки ледяных клиньев, складки, образующиеся при попеременном замерзании и оттаивании осадочных образований, солифлюкция, каменные россыпи и аналогичные отложения, некоторые из которых могут свидетельствовать о наличии в то время вечной мерзлоты. Наличие обширных песчаных дюн и лессовых отложений может служить указанием на развитие флювиогляциальных равнин, на сильные ветры и на отсутствие лесного покрова. Присутствие тундровой зоны в низах позднеледниковой колонки, составленной по данным пыльцевого анализа, указывает на холодный климат в центральной Европе. Фауна моллюсков и млекопитающих животных подтверждает эти данные.

Позер и Бюдель, в отдельности, пробовали наносить на карту климатические и растительные зоны плейстоцена. Позер нанес южную границу вечной мерзлоты согласно с распространением слепков ледяных клиньев и складок, образующихся в осадочной толще при замерзании, а северную границу леса—по данным пыльцевого анализа собранных Фербасом. Эти две границы были затем использованы для того, чтобы обрисовать границы нескольких растительных и климатических провинций. Добавочное изучение позднеледниковых песчаных дюн послужило основанием для составления карты Европы, отображающей среднее летнее атмосферное давление.

Бюдель поместил плейстоценовую границу морозной зоны соответственно с ее возможной зависимостью от положения снеговой границы на высоте 1000 м над уровнем моря, которую можно проследить в горах центральной Европы. Представления Бюделя отличаются от данных Позера, главным образом, касающихся Венгрии и Франции, где имеется недостаточно данных по пыльцевому анализу для выяснения характера растительности. Бюдель различает в Европе несколько растительных зон на основании распространения лесса и по данным пыльцевого анализа, если таковой производился. Бюдель считает, что ледяной покров во время плейстоцена имел лишь небольшое влияние на характер климатических зон Европы.

Хотя и имеются возражения по отношению ко многим данным использованным Позером и Бюделем для нанесения на карту растительных зон, эти карты служат основой для изучения воздушной циркуляции в плейстоцене.

Последовательность климатических изменений за последний холодный период большей частью определяется по стратиграфии перигляциальных лессов, переслаивающихся с почвами, и по их сопоставлению с последовательным развитием альпийского оледенения. Автор рассматривает противоречие в оценке вюрмского века, как единого (по Пеку), или как многократного (по Зергелю) оледенения. На основании недавних полевых работ и на основании данных углеродного метода,

собранных в Австрии и сопредельных территориях, автор высказывает сомнение о рациональности выделения межстадиального периода Геттвейга.

Падение средней годовой температуры в течении позднего плейстоцена, по данным понижения снеговой границы в горах, составило от 5 до 8°С, если считать, что возможный, вертикальный, температурный градиент атмосферы колеблется в пределах от 0,5 до 0,07°С на 100 м (скорость падения температуры). Расчеты, основанные на находках следов действия мороза и ископаемые растения в низменностях, дают основание предполагать, что снижение температуры составило от 10 до 12°С. По Мортенсону расхождение в числах является слествием различной скорости падения температуры в горах и в низменностях. Для того, чтобы произвести оценку этой гипотезы, автор рассматривает данные о современном изменении температуры в Арктике, субарктике и в субтропических областях.

Построение карт среднего атмосферного давления и характера воздушной циркуляции обычно основано на экстраполяции современных данных о снежных зимах, холодных летних периодах и других условиях, способствующих образованию оледенения. Это делается на основании предположения, что климат в плейстоцене отличался от современного климата другой периодичностью и другими путями движения воздушных масс. Возможные причины изменения в атмосферной циркуляции до сего времени являются весьма проблематичными, хотя с недавних пор больше внимания уделяется изменению солнечной радиации, в зависимости от изменения солнечных пятен.

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INTRODUCTION

During the cold phases of the Pleistocene, glaciers covered much of northern Europe and the Alps. The climate of the intervening area, generally called the periglacial region, was colder than the present, but the details of the climatic relations and the sequence of climatic changes depend on careful analysis of the geologic and biogeographic evidence and on the application of climatologic theory. The geologic features useful in such studies include principally frost-disturbed soils and wind-formed deposits. Biogeographic relations involve pollen analyses and the interpretation of mammalian and molluscan faunas. On the basis of these studies, maps have been drawn to show the major vegetational and climatic regions of Europe. The maps of climatic zones in turn lead to maps of Pleistocene atmospheric-pressure systems. Finally, the paleoclimatic maps and curves provide the basis of discussion of the possible causes of Pleistocene climatic change.

It is the purpose of the present paper to examine critically the several steps in this pyramid of compilation. Consideration is confined almost entirely to the last major cold phase of the Pleistocene (Würm, Weichsel), because so much more information is available concerning this epoch than for the rest of the Pleistocene. Eastern Europe is not included in much of the discussion.

The writer is obliged to the following persons for arranging excursions to important Pleistocene localities in Europe so that he might have at least some familiarity with the field relations: Karl Adam, Alfred Dücker, Julius Fink, Hans Graul, Karl Gripp, H. Jäckli, Sheldon Judson, Réné Tavernier, Fritz Weidenbach, and Ludwig Weinberger. He is also grateful to Ward Barrett, P. E. Cloud, C. S. Denny, S. S. Goldich, John T. Hack, J. Hoover Mackin, and Paul Woldstedt for critically reading all or parts of drafts of this paper, and to H. T. Mantis for discussion of some of the climatological problems.

GEOLOGIC RELATIONS

The Periglacial Region

Introduction. The term "periglacial" was introduced by the Polish geographer W. von Lozinski (1912, p. 1039) to emphasize the concept that central Europe during the cold phases of the Pleistocene was subjected to very severe climatic conditions marked by intensive frost action and wind action occasioned by the presence of the Scandinavian ice sheet to the north and the Alpine ice cap to the south. Various manifestations of a periglacial climate had long been known, and the history of this concept has been reviewed by Zeuner (1945, Ch. 1), Troll (1947), and Smith (1949).

Long ago Fisher (1866) attributed the unconsolidated debris known as "head" or "warp" in parts of England to Pleistocene frost action, and Blanckenhorn (1896) did the same for the "pseudo-moraines" in Germany. After Lozinski's summary appeared, interest in periglacial morphology increased, and observations were recorded by many writers from localities all over those portions of Europe not covered by the last ice sheet. Soergel (1919) and others showed the relation of the loess of central Europe to Pleistocene frost features and climate. During the same period, modern frost features were being studied and classified, impelled particularly by the work of Bertil Högbom (1914). All these studies led to the preparation of a book by Kessler (1925) on the Pleistocene climate and its geologic effects in unglaciated regions.

Some of the enthusiasm for the periglacial was opposed by the influential German geomorphologist W. Penck (1924) and others, who ascribed block fields and such "periglacial features" to modern processes. Soergel (1936) continued to emphasize the effects of climate on surface features through his work on fossil ice wedges, and since 1940 a new generation of workers have refined the concepts of frost soils and periglacial climate. Troll (1944) prepared a comprehensive analysis of the relationship of modern frost features to climate in different parts of the world. He followed it with papers devoted more specifically to the geomorphic processes active in the periglacial environment (Troll, 1947; 1948) and prepared a map of the Old World showing the distribution of modern frost soils with respect to altitude. Büdel (1951; 1953) and Poser (1948; 1950) prepared maps of Pleistocene vegetational and climatic zones of Europe, and Frenzel and Troll (1952) extended the surveys to Asia.

This more recent work of the German geographers has been supplemented by observations from France (Tricart, 1956a; 1956b), and from Poland, where the tradition of von Lozinski is revived by the work of Dylik (1956), Jahn (1956), and others with the publication of a journal devoted exclusively to the periglacial.

Investigations in Germany continue at a high rate (*e.g.*, Klimaszewski, 1959). In the United States, studies have not progressed far beyond the stage of infancy assigned them in 1949 by H. T. U. Smith, who prepared an excellent summary of the status of both the European and American work up to that time. The small volume of observations on periglacial features may not be a result solely of American disinterest in this field, for it is probable that the periglacial zone in this country was much smaller and less pronounced than it was in Europe. At any rate, the possibilities for a broad synthesis in this country such as has been prepared for Europe are greatly restricted by the paucity of data on Pleistocene climate and vegetation.

The review by Smith (1949) appeared when the summary studies in Europe were just being started. So many amplifications, refinements, and paleoclimatic maps have been presented in the last decade that a more detailed discussion of this later work is here presented.

Periglacial processes. Troll (1948, p. 13 ff.) identifies the following processes as being dominant in the periglacial environment:

(1) Congelifraction—breakage of rocks by frost action
(2) Congeliturbation (cryoturbation)—disturbance of the ground by frost action, as in the formation of stone nets
(3) Solifluction—flow of saturated soil on slopes in cold regions
(4) River-gravel deposition—caused by excessive supply of rubble by congelifraction and solifluction
(5) Wind deflation (gelideflation) and deposition
(6) Cryoplanation—final smoothing of the terrain as an end product of the activity of all these processes.

Troll (1947; 1948) proposes that the periglacial (subnival, solifluidal, soligelid) cycle of denudation should take its place along with the fluvial, marine, eolian, glacial, and karst cycles as a major geomorphic phenomenon, for it now affects a large part of the continental surface and during the Pleistocene it affected an additional area to such an extent that fluvial processes have subsequently produced very little modification. A triangular diagram (Fig. 1) designed by Troll emphasizes the relations among humid, arid, and "nival" climatic regions.

Büdel (1953, p. 251) has questioned the propriety of the term *periglacial* for such environments, which are not necessarily anywhere near a glacier. Actually, Lozinski (1912) had introduced this term to substitute for the word *subglacial*, which Andersson (1906) had used for ancient solifluction deposits in the Falkland Islands and which today is used in a completely different sense. It has generally been assumed that the severe frost action and other indicators of cold climate in central Europe in the Pleistocene were a direct effect of the

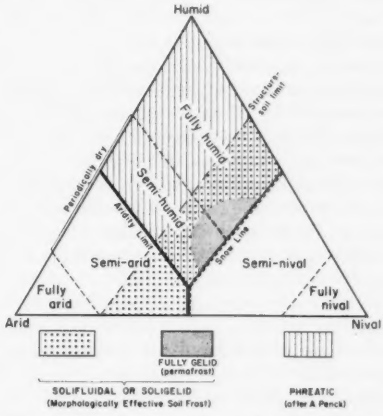

Figure 1. Climatic regions of the world according to their physiographic effects, expressed diagrammatically (Troll, 1948, Fig. 1)

proximity of the expanded north European and Alpine ice sheets, and that the environment is therefore "periglacial." Büdel, however, believes that the basic climatic change of the Pleistocene involved the general shifting of climatic zones toward the equator, that the frost soils and related cold-climate features were a direct manifestation of this shift regardless of glaciation, and that the presence of the ice sheets (themselves of course also a product of climatic change) only locally served to intensify the cold climate.

Evidence for Büdel's point is difficult to adduce, although it is not hard to show that the *retreating* ice sheet had only a limited effect on the climate of the surrounding region. Büdel, for example, emphasizes that the frost features in question are largely confined to the area *beyond* the deposits of the last glaciation in northern Europe, and that these deposits themselves do not in general bear such features.

During the approximate 10,000 years of ice retreat from the outermost Weichsel moraine (Brandenburg) in northern Germany to the Central Swedish moraine, the climate was sufficiently mild that no extensive frost features were formed on the deglaciated terrain (Johnsson, 1958, p. 338). Although some frost action may have been important during the early stages of retreat, the terrain was probably stabilized with vegetation shortly after the ice started its retreat across the Baltic Sea from the Pomeranian moraine; organic sediments began to accumulate in lakes, and the pollen content shows that forest began to invade the tundra as early as *ca.* 13,000 years ago (Bölling zone). Even though two later phases of forest recession are recorded in pollen diagrams (Older and Younger Tundra zones) there is little evidence for frost-soil formation in southern Sweden, which should technically have a periglacial climate because the ice sheet was so close. The climatic amelioration for the late-glacial was thus rapid. It may have been even more rapid than the changes recorded either by moraines or by pollen sequences, because of retardation factors, *i.e.* the time required to melt the great mass of ice stored from the earlier climatic regime, and the time required for invasion and succession of plants on bare terrain.

On the other hand, involutions (Poser, 1948, p. 56) and ice-wedge casts (Johnsson, 1958) have been found *under* drift of the Weichsel ice advance, implying that the climatic change that brought down the ice sheets had already produced a belt of permafrost. Such buried permafrost may then have thawed under the insulating cover of glacier ice and may never have formed during or after ice retreat. Büdel, therefore, believes that the Pleistocene "periglacial" climatic zones of Europe were a direct effect of world-wide cooling and not primarily a secondary effect of glacier expansion. He thus holds that the term *periglacial*, as applied to the Pleistocene environment of central Europe, is misleading. He now uses the term only with quotation marks for this situation, and deplores its use for alpine or other environments clearly far from the ice.

Unfortunately, there seems to be no single satisfactory substitute term. Troll (1947, p. 169) has discussed the terminology particularly with respect to alpine environments and suggests that the term *solifluidal* (or *soli-gelid*) be applied to those areas characterized by frost-soil formation and solifluction. He also uses *subnival* for such a zone *below* the snow line (not *under* the snow; *cf.* subglacial, *under* the ice). Perhaps straightforward terms like *tundra zone* or *frost-soil zone* or *permafrost zone* would be more satisfactory, depending upon the nature of the criteria by which the zone is identified. In the present review these designations will be used where appropriate, but the generalized term *periglacial* will be retained for the Pleistocene relations to describe in a broad sense the middle-latitude zone of that time characterized by extensive frost action, wind action, and/or tundra vegetation, with the understanding that the climatic characteristics of this zone might not necessarily arise from the presence of the ice itself. Modern polar, subpolar, and alpine zones may have comparable environments.

Frost soils. Whenever frost action occurs in the soil, some disturbance may be expected. In temperate climates such disturbances are not sufficiently frequent to permit a recognizable pattern to develop, or the change is masked by other processes that occur during the balance of the year. In the high mountains of the tropics, however, where there is nightly frost but no annual temperature cycle, repeated needle-ice formation can produce miniature stone nets, stone stripes, etc., a few inches broad (Troll, 1944, Ch. 5, 7). Permafrost is clearly not a requisite for the miniature forms, and in fact is not a requisite for such moderate-sized stone rings as are found on favorable sites in certain subpolar areas (Hopkins *et al.*, 1955, p. 138). The very large boulder rings several meters across, however, are probably restricted to permafrost areas, and the frost heaving of such large stones must be favored by the presence of a permafrost layer beneath that provides not only a cold reservoir for freezing of the active layer upward from below (Taylor, 1956, Ph.D thesis, Univ. Minn., p. 110) but also an impermeable barrier to downward infiltration of thaw water.

Frost soils have been classified in numerous ways, and there are as many terms as there are stones in a stone ring—and almost as many hypotheses for the formation of the different features. Recent summaries of all or some types of frost soil have been prepared by Troll (1944), Cailleux and Taylor (1954), and Washburn (1956). For our purposes, in which we wish to identify areas of Pleistocene permafrost as well as areas where frost but not permafrost was present, it is useful to emphasize certain types.

One group of frost features with surface expression is termed patterned ground (Washburn, 1956). We may consider this in three

categories: (1) ice-wedge polygons and related fracture features, generally confined to permafrost; (2) stone nets, stone stripes, and related forms in which there is internal sorting of particles in the soil but for which permafrost is not a necessary prerequisite; and (3) other patterns that involve only vegetation or unsorted fine soils (peat rings, earth hummocks, vegetation stripes); these are rarely preserved from the Pleistocene and will not be further considered.

Involutions, another type of frost feature, are formed beneath the surface by freezing of the summer thaw layer of permafrost. They may have no surface expression but are easily preserved (although not so easily discovered). This type is believed to require permafrost generally for its formation and so is particularly useful in reconstruction of Pleistocene climate.

A third group of frost features may be assigned to the process of solifluction, which involves downslope movements of water-saturated debris (Andersson, 1906). The water saturation in solifluction materials comes about generally not only through infiltration of snowmelt, as Andersson thought, but also through ice segregation during the freezing cycle (Taber, 1943, p. 1457). If a base of frozen ground (permanent or long-seasonal) is present it prevents infiltration of thaw water and provides a gliding surface as well. Solifluction is thus favored in regions of permafrost or long-lasting seasonal frost, but may occur in fine-grained debris in front of springs or other localized sources of moisture. In consideration for its original definition by Andersson, it seems best to confine the term to those situations in which frost has at least some part to play.

As far as climatic reconstructions are concerned, the location of the Pleistocene permafrost boundary is critical, because this line gives a clue to the mean annual temperature of the region (Black, 1954, p. 843), although the clue may not be very precise (Brown, 1960, p. 171 ff.). Therefore, frost features that require permafrost for their formation should be distinguished from those that do not. Thus Troll (1947, p. 165) distinguishes three zones: (1) polar zone of permafrost, (2) subpolar (and alpine) zone of long-lasting seasonal frost, and (3) tropical high-mountain zone of daily frost (also found on certain high-latitude oceanic areas like Iceland).

As a matter of fact, the modern permafrost limit is so difficult to map that three zones are often distinguished. Black's (1954, p. 841) map of modern permafrost shows zones of *discontinuous* and *sporadic* permafrost 500–1000 miles broad south of *continuous* permafrost. The occurrence of permafrost depends on a variety of geologic and hydrologic requirements. The mean annual temperature required for permafrost is variously estimated to be below −5°C (Mortensen, 1952, p. 153) or below 0°C (for sporadic permafrost, Black, 1954, p. 839); in addition, other climatic elements such as continentality also affect its distribution (Brown, 1960).

Numerous frost features are described or mentioned below, with special comment concerning the preservability of the features after a climatic change, and thus their utility as indicators of past permafrost or of specific past climatic characteristics. Many frost features are not mentioned because they are not common or do not have a particular value in Pleistocene studies.

ICE-WEDGE POLYGONS: Ice-wedge polygons are found today in regions of continuous permafrost in favorable terrain. They were first made famous by the work of Leffingwell (1919, p. 179–243) on the coastal plain of northern Alaska, and the mechanics of their formation have been more recently discussed by Black (1954, p. 846).

Although Taber (1943, p. 1524) and Dücker (1951) have made a case for the origin of ice wedges merely by ice segregation in fine-grained soils, the prevailing theory for their formation (Leffingwell, 1919, p. 205; Black, 1954, p. 844) calls upon ground contraction resulting from severe winter cold. In winter the wedge-shaped contraction cracks become filled with snow and hoar frost and in the summer by meltwater and slump material. If the ice wedge does not melt out in summer it forms a plane of weakness for cracking the following year, and the feature may thus widen. The width of the wedge thus reflects the frequency of cracking, and depth reflects the severity of the winter cold. Secondary factors are the materials, moisture content, topography, etc.

Repeated contraction and expansion produces deformation structures in the ice crystals within the wedge (Black, 1954, p. 845) and in the material of the adjoining walls. The walls may thus be pushed up into a pair of narrow ridges bounding the wedge, and the resultant land forms in many regions consist of a polygonal network of narrow troughs, locally marked by long ponds; in other areas larger lakes or bogs

occupy the low areas in the centers of polygons (Hopkins *et al.*, 1955, p. 138).

Ice-wedge casts of Pleistocene age have long been identified in Europe as a record of Pleistocene permafrost (Soergel, 1936). They are recognized in the field by the wedge form, the sharp discordance of unsorted debris against the wall, the vertical orientation of elongate stones, and the distortion of the layers adjacent to the wedge. The filling generally consists of heterogeneous material that sloughs off the walls or is washed in or blown in at the top when the ice melts. In many places the wedge filling and the adjacent ground are truncated by a solifluction deposit in which the postglacial soil is developed. Such overburden may obscure the polygonal pattern in plan view, and wedges are commonly observable only in cross-section exposures. Miniature ice-wedge fillings only a few inches deep are found in deposits of loess reworked by solifluction. These do not require permafrost for their formation, for simple single winter freezing might produce the requisite contraction.

Certain other soil structures may be confused with ice-wedge casts, and it is necessary to observe all or most of the criteria listed above to be certain of the genesis of the feature (Johnsson, 1959). For example, soil pendants have a similar form in cross section but are conical and are isolated in distribution rather than occurring in a polygonal network (Yehle, 1954). Soil pendants are particularly well developed in calcareous gravels. They consist of the localized downward extension of the B horizon of the soil and represent spots in which the descending weathering solutions become concentrated. Unlike wedge casts, these features do not show oriented stones or deformation of the walls. Polygonal wedge patterns can also be formed by alternate cracking and swelling of the soil of desert playas as a result of wetting and drying (Knechtel, 1952, p. 694–698). Other desert polygonal patterns reflect lateral expansion resulting from growth of salt crystals (Bobek, 1959, p. 22). The gilgai soils of Australia result from swelling of clayey subsoils; furthermore, they are not restricted to desert regions but are also found in temperate forest areas (Hallsworth, 1955). Some of the so-called mima mounds of western United States form a polygonal pattern that has been attributed, most recently by Péwé (1948), to Pleistocene frost action, but other authors defend origins unrelated to frost processes. Polygonal patterns on the floor of Glacial Lake Agassiz in North Dakota have been ascribed to periglacial frost cracking (Horberg, 1951), but other explanations have also been proposed (Nikiforoff, 1952; Mollard, 1957).

Even where properly identified as ice-wedge casts the structures may not record continuous permafrost, for Hopkins *et al.* (1955, p. 138) found modern ice-wedge polygons in Alaska in regions of discontinuous permafrost.

STONE RINGS AND RELATED FEATURES: Stone rings, nets, garlands, stripes, and other sorted frost features are perhaps the best-known manifestation of extensive frost action and are found widely distributed in suitable sites in polar, subpolar, and alpine regions. Miniature forms may be found on tropical high mountains and in certain regions with oceanic climate (*e.g.*, Iceland). Extinct forms have been reported from a great many areas in Europe and elsewhere; they indicate the distribution of past intensified frost action probably under conditions of sparse cover of vegetation. Although it is commonly stated that such features require permafrost for their formation, Troll (1944, p. 566) points out that the miniature forms produced by daily frost alternation do not involve permafrost, and other forms may result from a condition of long-lasting seasonal frost rather than permafrost. Recent studies in Alaska by Hopkins *et al.* (1955, p. 138) show that stone polygons may be formed in areas without permafrost, although they are more common where permafrost is present.

Stone rings and related features, where stabilized by vegetation cover and modified by chemical weathering, suggest past conditions of colder climate and presumed absence of forest cover. Probably only the larger features require permafrost for their formation. When covered with heavy vegetation such features are rarely detected, and they are often so shallow that even in vertical cuts any patterns have been obscured by later soil formation. Occasionally, however, one can sight local concentrations of vertically oriented stones, an almost certain clue to upward frost heaving of individual stones. Sometimes these may be related to weak involution patterns or to ice-wedge structures, but in other areas they may record the remnants of a stone ring or other sorted feature.

INVOLUTIONS: Involutions are tightly folded laminae of silt and fine sand generally believed to be produced near the base of the "active layer" of permafrost during fall freezing. The recognition of these features in the surficial deposits of Europe as records of Pleistocene peri-

glacial climate dates from the work of Steeger in 1925 (Steeger, 1944). They are generally called in German *Würgeboden* (literally strangle soil) or *Taschenboden* (pocket soil). They have also been termed *Brodelboden* (bubble soil), but this implies acceptance of the hypothesis that the features are formed by convection in supersaturated soil during the freezing process, as thought by Gripp (1927), who introduced the term. The word "involution" was introduced by Sharp (1942), at the suggestion of Kirk Bryan, and has been widely accepted as a descriptive term noncommittal as to the exact mechanism of formation. The French equivalent is *plication*.

Involutions consist of highly crumpled laminae with balloonlike upfolds or downfolds that in cases are pinched off at the neck to form isolated pockets. The vertical axes of the structures indicate that the lateral pressure was uniform, and in this respect they may differ from crumpled beds produced by subaqueous slump, dragging of icebergs, solifluction, or comparable processes that involve a shearing force. The pressure is believed to be caused in the autumn during final inhomogeneous freezing of the lower part of the active layer of permafrost, *i.e.*, the surficial layers of annual freeze and thaw (Bahr, 1932, p. 29 ff.). The active layer freezes from the surface downward and to a lesser extent from the base up (Taylor, 1956, Ph.D. thesis, Univ. Minn., p. 110). Although the permeable laminae may be dehydrated in the process of segregation of ice into freezing centers, the finer-grained laminae retain water and become mobile under the freezing pressures from above, below, and within. The contained water cannot infiltrate downward because of the impervious character of the underlying frozen ground. Expansion downward is inhibited by the resistance of the permafrost, and expansion upward by the thickening crust. The deformation is therefore confined to the lower part of the active layer.

The requirements of permafrost below the involution zone are inferred from the belief that involutions do not form in cool-temperate regions, which may have deep winter freezing but do not have permafrost. Soil distortions similar to involutions have been studied in the process of formation in Alaska (Hopkins and Sigafoos, 1950, p. 98), where they are related to differential freezing of shallow vegetation polygons that have developed in a permafrost region from frost scars in a once-continuous vegetation mat. Certainly, however, the presence of permafrost may not be essential where poor drainage, growth of vegetation tussocks, and other conditions may be favorable for differential freezing and ice segregation. In Massachusetts, where no permafrost is now present, modern soil deformation resulting from frost action around tussocks has produced structures suggestive of involutions (Sigafoos and Hopkins, 1951). The writer has recently studied exposures in a shallow depression near North Branch in eastern Minnesota in which an involution layer is present that is typical of the form so widely described from the permafrost region of Europe (Fries, Wright, and Rubin, 1961). The organic material incorporated into the involutions, however, is 2520 years old, according to a radiocarbon date, so the feature is much too young to permit the assumption of permafrost. The mean annual temperature in the region at present is about 6°C.

SOLIFLUCTION FEATURES, ASYMMETRIC VALLEYS: Troll (1947, p. 166) has suggested that the term solifluction be broadened to include internal movement of soil particles as in stone rings (microsolifluction) as well as downslope movement (macrosolifluction). This internal movement, which may result in the formation of patterned ground, is included in the terms cryoturbation of Edelman *et al.* (1936) or congeliturbation of Bryan (1946). Troll's concept of solifluction is expressed in Table 1, which shows different bases of classification.

For our purposes we shall not include in the term solifluction any internal movements (microsolifluction of Troll) unless some slope movement is involved as well (as in stone stripes).

Some solifluction deposits have distinctive form or surface expression as well as content, and where preserved may be useful as paleoclimatic indicators. On slopes they may take the form of lobes or steps or stripes, in which case some sorting of particles may be involved. More often they have no surface micro-expression but impart smooth contours to the terrain by inhibiting the development of stream gullies.

Solifluction may supply streams with so much debris that extensive gravel deposition results; river terraces in the Pleistocene periglacial area of Europe may have this origin.

Solifluction materials usually consist of locally derived angular rock debris unsorted and unstratified in a matrix of fine particles. As such they may be confused with local till or landslide debris, and inference as to the role of frost is often difficult when the solifluction debris

may be derived from upslope till and may thus include erratic and nonangular rock fragments. Stone-orientation analysis may be used to relate solifluction deposits to the slopes on which they occur (Lundqvist, 1948), and stone-roundness analyses have been used for a similar purpose (Cailleux, 1951).

The very upper ends of valleys in regions of present or past intensive frost action are marked by smooth slopes and gentle trough-like cross sections as a result of movement of saturated surface soil. Exposures show thick accumulation of debris at the base of the side slopes and along the valley floors. Such small smooth valleys in central Germany are called *Dellen* (dells) or *Muldentälchen*. Their slopes are now stabilized, and their floors dry, for the forces of solifluction which are believed to have molded them are no longer active. Even at the time of formation there was probably little stream action, as is indicated by the nature of the deposits in the valley bottoms.

Some solifluction valleys show a distinct asymmetry in cross section. Asymmetric valleys have been noted in many parts of the world and have been attributed to a variety of factors. Although some asymmetric valleys are certainly controlled by rock structure, regional tilting, or possibly the deflection of running water by the Coriolis force, most types are attributed to some kind of climatic control. Those found in central and western Europe are generally steeper on the south- or west-facing slopes. Poser (1948, p. 54; Poser and Müller, 1951) attributes the asymmetry to greater insolation on these slopes. Early summer melting of permafrost on these exposures allows the stream to erode them more rapidly and thereby steepen them. The migration of the stream to the steeper slope is aided by long-lasting solifluction on the gentle slope, as the ground slowly thaws during the summer; the accumulating solifluction debris at the base of the gentle slope helps to push the stream to the opposite side. In many cases the valleys are now dry; it is reasoned that the runoff now infiltrates, whereas during the cold period the permafrost prevented infiltration.

Other workers, notably Büdel (1953, p. 255), attribute valley asymmetry in this region to preferred accumulation of loess and drifting snow on the east-facing (lee) slopes by westerly winds. In the summer the mixture of loess and snowmelt flows as a solifluction sheet down the slope over permafrost and crowds the river to the opposite (west-facing) slope, which generally has no mantle of solifluction debris.

The relative importance of insolation or wind in the formation of periglacial asymmetric valleys is difficult to evaluate. Troll (1948, p. 18) suggests that a combination of the two factors may produce the observed relations.

Excellent examples of valley asymmetry can be found in the terrace landscape of the Alpine foreland of southern Germany and adjacent Austria, where the higher Pleistocene terraces have all been eroded on the east sides of the valleys and preserved on the west sides, because the axial rivers migrated eastward during successive cold phases of the Pleistocene (Büdel, 1944, p. 494 ff).

The dells at the very heads of the valleys, with their solifluction fill on rounded bottoms and with no axial streams, are generally symmetrical, but Poser (1948, p. 54; Poser and Müller, 1951) describes some that have steep slopes facing east or north, in contrast to the sense of asymmetry downstream. Poser reasons

TABLE 1. FORMS OF SOLIFLUCTION
(Troll, 1947, p. 168)

Basis of classification	Types of solifluction	
Form	Differential solifluction (with sorting)	Amorphous solifluction (no sorting)
Terrain and movement	Macrosolifluction (downslope)	Microsolifluction (radial)
Time interval	Seasonal solifluction	Daily or short-term
Kind of ice	Frozen-ground solifluction (seasonal or perennially frozen ground beneath solifluction layer)	Needle-ice solifluction (substratum not frozen)

that with no axial stream the sunny slopes are not subject to undercutting, and apparently that more rapid flattening is caused in the spring by solifluction on the sunny slope rather than by summer solifluction on the shaded slope. Analogy is made with the asymmetric valleys described in the permafrost region of Siberia (Schostakovitch, 1927, p. 420), which have steep north-facing slopes. It should be noted, however, that the widely quoted Russian author did little more than represent the valleys in a diagram and did not explain their genesis.

Most of the climatically controlled asymmetrical valleys described in Europe appear to have differential accumulation of solifluction deposits on opposite slopes. A recent soils study in the Chiltern Hills of southern England, however, showed that asymmetry of dry valleys occurs in the bedrock itself, with the solifluction deposit being thin and irregular (Ollier and Thomassin, 1957). Here we have an instance of river migration during downcutting, rather than lateral crowding by differential deposition. The authors still reason, however, that differential insolation must affect the slope creep and thus the rapidity of slope retreat. The gentle slope is interpreted as simply the slipoff slope left by the stream as it cut downward and laterally.

Alexandre (1958, p. 264 ff.) found valleys in the Ardennes that had been modified in two ways—a solifluction phase in the upstream segment, in many cases with asymmetry, and a rainwash phase downstream, marked by gullying by the axial stream. The junction between the two zones shifted up- and down-stream during the Pleistocene, as recorded by combination cross sections—the upper part of the section may show an asymmetrical solifluction slope, and the lower part a gullied slope. Alexandre attributes the solifluction phase to the time of permafrost and the gullying phase to the time of maximum snowfall and runoff (time of maximum glaciation in northern Europe).

Most authors feel that solifluction valleys and in particular asymmetric valleys reflect a periglacial climate, but Ollier and Thomassin (1957) admit that there is no proof that some of the valleys in question in western Europe may not reflect interglacial conditions rather than periglacial. This disarming statement causes one to be skeptical of immediate acceptance of asymmetric valleys as indicators of permafrost. Soil creep apparently flattens slopes, at least until some equilibrium condition is reached, and differential exposure to the sun (or some other climatic control) may result in a different equilibrium angle on opposite slopes.

Regardless of the details of the causes for climatically controlled asymmetry in small valleys, it has been assumed that the features indicate the existence of permafrost at the time of formation (Poser, 1948, p. 54; Troll, 1948, p. 18). This assumption is based on the fact that extensive solifluction requires supersaturation of soil on gentle slopes, and such a condition is favored by the presence of frozen ground that prevents infiltration of the thaw water from the active layer. Actually, however, this impermeable substratum may merely be the lower part of the layer of seasonal frost, which may not thaw until very late in the season after several months of solifluction. Certainly solifluction deposits can be found forming today in regions without permafrost. Poser states that such asymmetrical valleys must be criteria of ancient permafrost because similar valleys are found forming today only in regions of modern permafrost (*i.e.*, Siberia). In view of the fact that only this one Siberian study of modern asymmetrical valleys can be cited and in view of the variable factors and hypotheses for their formation, those asymmetric valleys involving differential solifluction should not be emphasized as reliable criteria for past permafrost, but merely as criteria for a past cold climate.

Soil creep, of course, is not confined to permafrost regions and in fact is not confined to areas of frost action (solifluction). The burden of proof must rest with each investigator to demonstrate that frost action and particularly permafrost are involved in the soil creep. This may be easier to do if creep deposits in the valley bottoms are closely associated with periglacial loess or other deposits that are themselves demonstrably Pleistocene. Measurements of particle shape, sorting, and other characteristics of these deposits have been made by Cailleux (1951) in efforts to identify diagnostic criteria.

BLOCK FIELDS, TALUS, ROCK GLACIERS: Block fields and block streams may be produced by frost breakage of rock outcrops accompanied by gradual downslope migration of the fragments. Where later stabilized by vegetation, soil formation, or bog development they may be preserved with little modification. Because block fields readily form today in polar, subpolar, and alpine regions, the stabilized block fields of

central Europe and similar temperate regions are generally attributed to Pleistocene climatic conditions, although the cessation of block-field development may locally result from factors other than climatic change. Büdel (1944, p. 487), who proved the antiquity of some block fields in the Erzgebirge by the identification of the entire postglacial pollen sequence in an overlying bog, studied the distribution of stabilized block fields in mountains of central Europe in an early effort to determine the altitudinal limits of Pleistocene frost climate.

Rock accumulations on steep slopes below bare cliffs form readily in regions with frost weathering. In alpine regions well above the tree line the talus cones below cliffs give way to rock glaciers, which extend along the valley floor much like true glaciers. Measurements of rock glaciers in the Alaska Range imply that the thickness of a rock glacier, and presumably of the source talus, must generally be at least 100 feet before movement can take place (Wahrhaftig and Cox, 1959, p. 426). Where stabilized and no longer apparently forming, these features are attributed by many to past climates more severe than the present. Forested talus accumulations on slopes in the mountains of central Germany and France may fall in such a category, as do beds of angular rubble derived from roofs of limestone caves and rock glaciers. The paleoclimatic significance of these features is difficult to evaluate, and each occurrence must be considered on its merits. Local conditions unrelated to climate may cause a given talus to become stabilized—perhaps a change in the extent of exposure of the cliff that serves as the source. As far as caves are concerned, however, Movius (1960, p. 380, 384) points to impressive evidence in the caves of southern France that the roof-rock debris was produced by frost action: the Upper Paleolithic culture layers, dated by C^{14} in the range 30,000 to 13,000 years ago, are filled with rock chips, whereas the post-Paleolithic contain few. It can be safely inferred that the process responsible for loosening the roof rock is not so active now as it was during the late Würm.

DISTRIBUTION OF MODERN FROST SOILS: Troll's (1944; 1948) studies of the distribution of modern structure soils (patterned ground) and solifluction deposits with relation to latitude over the various climatic zones of the world have given a broad basis for his presentation of a map of the Old World showing the altitudinal limit of these features (Troll, 1947, Fig. 1). Troll's map shows that the limit of frost soils bears a close relation to the snow line but is many hundreds of feet lower. Thus it rises from sea level in the polar regions toward the equator, reaching an elevation of about 5000 feet in the mountains of central Germany, 6500 feet in the Alps, and 11,000 feet in the Atlas Mountains. The line reachest its highest position in the dry mountains of the subtropics (about 15,000 feet in Tibet and the Andes) and is lower in the more humid tropics of East Africa (13,500 feet), although the observations in low latitudes are scattered. The effect of precipitation on the climatic structure–soil boundary can be seen in local areas, *e.g.*, a rise from 6000 feet on the outer Alps to over 7000 feet in the drier central Alps, or a rise from 9000 feet in the Caucasus and West Pamir to 15,000 feet in the drier East Pamir. So also goes the present snow line, at higher elevations (Klute, 1928).

Wind action. The importance of wind action in the Pleistocene periglacial regions of Europe has been realized for many years and has recently been discussed by Smith (1949).

SAND DUNES: Morphology of sand dunes may reveal the direction of dune-forming winds and thereby some of the characteristics of Pleistocene climate. An early comprehensive study on this problem was presented by Ivar Högbom (1923), who inferred westerly winds for northern and central Europe on the basis of dune form and structure. Enquist (1932), however, inferred easterly winds from the same data, and this conclusion was used in support of the famous Hobbs theory of the glacial anticyclonic wind circulation in the periglacial region. More recently Poser (1950) prepared a map of late Pleistocene pressure patterns and wind systems on the basis of dune form in northern and eastern Europe. This work is discussed in the section concerned with Poser's reconstruction of Pleistocene climatic zones.

SAND GRAINS AND VENTIFACTS: Cailleux (1942) considered that wind action was so prevalent in the Pleistocene periglacial regions that individual sand grains carried by the wind acquired a rounded form and pitted (frosted) surface—features that were retained even though the sand grains were incorporated in other deposits like till, glaciofluvial sediments, frost soils, etc. This assumption was based on his study of about 3000 samples of Pleistocene deposits over Europe, in which he determined the percentage of frosted grains in the coarse sand fraction. A map (Cailleux, 1942, Fig. 16, or Wright, 1946, Fig. 1) shows that the belt of high percentage of frosted sand grains lies just

outside the drift border for the last glacial phase of the Pleistocene in northern Europe, and that the highest values are in the eastern part of this belt in Poland (80–100 per cent frosted grains), where the continental climate apparently permitted more extensive wind abrasion. Cailleux's study has been discussed at length by Wright (1946). The method has been recently applied in less elaborate fashion in the United States, but with no results concerning Pleistocene climate (Schneider and Cailleux, 1959).

The association of ventifacts with frost soils has been observed in many Pleistocene periglacial regions (Smith, 1949), but the only regional study has been that of Cailleux (1942), who noted the more common occurrences of large multifaceted ventifacts in northern and eastern Europe in the area of frost soils and frosted sand grains. He reasoned that in these regions such stones (larger than 10 cm) could be overturned by frost action and thus cut by wind abrasion on more than one side.

LOESS: The eolian origin for the upland loess of central and western Europe is now almost universally accepted. In many areas each layer of homogeneous loess is underlain by a layer of solifluction loess that includes soil and other materials transported downslope. Three climatic phases are generally recognized for each loess layer: (1) a cold and moist phase involving some loess deposition but accompanied by much soil erosion and solifluction, (2) a cold dry phase with dominant deposition of homogeneous loess, and (3) a warmer interstadial or interglacial phase with formation of soil *in situ* and without loess deposition.

Detailed stratigraphic studies, especially in Austria and southern Germany, have shown the relatively delicate regional climatic control on loess deposition and modification (Fink, 1956; Brunnacker, 1956). In the drier eastern portion of Austria and parts of adjacent Czechoslovakia the loess is little disturbed by solifluction or frost cracking, except in favorable localities. In the more humid regions of western Austria and southern Germany, however, the basal solifluction loess is more conspicuous, and in many areas most of the soil formed during the preceding climatic phase has been eroded and transported downslope by solifluction to be redeposited along with fresh loess or other materials.

The source of the loess is predominantly glacial meltwater plains or valley trains, which dry periodically to permit deflation of the silt. Studies have shown consistent changes in the direction away from such sources, such as decrease in thickness and grain size (Schönhals, 1953). To the extent that such sources are required, the formation of loess is thereby restricted and is a secondary result of glaciation. Weidenbach (1952), in emphasizing this view, points out that most of the loess deposits of central Europe were blown from the glacial meltwater trains of the Rhine and Danube rivers and their major tributaries, and that the loess of Belgium and northern France had a source in the coastal meltwater plains since submerged by the rise in sea level. He further stresses that each loess deposit in a stratigraphic succession may therefore be correlated with a glacial terrace and thence with a moraine. The problems of regional correlation and chronology thus become involved.

Others believe that much of the loess of periglacial Europe is derived instead from the barren terrain with sparse vegetation cover away from major meltwater valleys, particularly terrain that is subject to intensive frost action which not only produces silt-sized particles by frost breakage but keeps the soil stirred up enough to reduce the plant cover and to inhibit the formation of the kind of stone pavement that prevents deflation (Dücker, 1937).

A third view about loess genesis is expressed by Büdel (1953), who considers that loess formation requires not only a suitable source region, whether meltwater plains or frost-stirred uplands, but also suitable conditions for accumulation. He points out that loess does not occur in northern Europe immediately adjacent to the drift border of the last glaciation, but is found instead in regions somewhat farther from the ice front. It might be suggested that lack of loess close to the ice is rather a reflection of the fact that glaciofluvial sources are sandy rather than silty in this province and that sand dunes are more common than loess deposits, at least in northern Europe. Büdel, on the other hand, points out that loess is absent also at higher elevations in the German Mittelgebirge. He believes that frost action close to the ice sheet and at high elevations was too intensive to permit the accumulation of loess—this is his frost-rubble tundra zone where the vegetation cover was sparse and where the loess falling on the surface was churned into the soil by frost processes. In regions where frost action was less pronounced, the loess would be entrapped by the heavier cover of grass or herbs and would not be so readily stirred into the soil beneath.

Hack (1953, p. 184) has taken issue with Büdel's requirement of tundra or grassland for loess deposition, pointing out that Péwé (1951) noted modern loess deposits in Alaskan forests near important glaciofluvial plains. If an adequate source is present there seems to be no reason why loess could not accumulate under forest as easily as under herbaceous vegetation. It must be admitted that the gross distribution of Pleistocene loess in Europe suggests that the continental interior (Lower Austria to Ukraine) provided the principal depositional sites, and here the forest cover was restricted, according to pollen analysis. The relations among vegetation, loess deposition, and source area might better be tested in the United States, where there are thick loess deposits in the Middle West but less evidence for the type of broad tundra zone that characterized most of Europe during the Würm.

BIOGEOGRAPHY

The shifting distributions of plants and animals during the Pleistocene provide a record of climatic change, and Deevey (1949) has discussed at length some of the ecological factors as well as the sequence as it was understood at the time.

Plant macrofossils are not commonly preserved in Pleistocene sediments, and where preserved they often record microclimatic environments like swamps that might not be particularly diagnostic of the regional vegetation. It may be mentioned, however, that the important Alleröd oscillation of the late glacial of Denmark was established on the basis of plant macrofossils (Hartz and Milthers, 1901), and that Nathorst (1914, p. 290 ff.) long ago emphasized the frigid character of the Pleistocene climate of central Europe on the basis of macrofossils of *Dryas* and other arctic plants in southeastern Europe. Iversen (1954) has discussed the value of macrofossils, especially of aquatic plants, in climatic reconstructions. Macrofossils have certain advantages over pollen grains in that specific rather than only generic determinations can often be made, and further that the fragments are too large to have been brought to the site of deposition from another environment far away (Godwin, 1956, p. 6–10). They rarely can be analyzed statistically like pollen grains, however.

Plant microfossils, especially pollen, have been much more useful in delineating regional vegetation because, although the pollen is commonly preserved only in lakes or bogs, it is blown to these sites from the surrounding upland terrain. Pollen occurs in sufficient quantities that it may be treated statistically—a modern detailed pollen analysis of a single sample involves the enumeration of 300–1000 grains, and thus the preparation of a pollen diagram through a pollen-bearing Pleistocene sediment may involve the counting of 20,000–100,000 grains. Numerous potential sources of error are involved in the method, such as differential pollen production and dissemination for different plant types, differential sedimentation and preservation of pollen grains, redeposition of older pollen grains, and other factors reviewed by Faegri and Iversen (1950, p. 90 ff.). Despite these difficulties the method has been remarkably successful in elucidating particularly the Late-glacial and Post-glacial vegetational succession of Europe, as seen from the almost identical diagrams from hundreds of localities (Firbas, 1949). As a quantitative or detailed representation of vegetation or ecological relations surrounding the site of deposition, the standard pollen diagram is somewhat less successful, for there is only a gross similarity between the modern vegetation and the modern pollen precipitation at those sites at which comparisons have been made (Faegri and Iversen, 1950, p. 82 ff.; Davis and Goodlett, 1960). Nonetheless, pollen analyses furnish by far the most abundant data for paleoclimatic reconstruction for continental regions.

The Late-glacial and especially the Post-glacial pollen sequence has been worked out with great precision for most of central and northern Europe (*e.g.*, Firbas, 1949, p. 294–346; Godwin, 1956, Ch. 3; Iversen, 1954). Correlation of the Late-glacial pollen zones with moraines is possible by the varve and radiocarbon chronologies. The Late-glacial diagrams record a tundra or at most a park-tundra landscape that reflects much lower temperatures than today. The absence of pollen in underlying sediments is generally interpreted as a result of climatic conditions so severe that the vegetation was indeed sparse, or that organic sediments were thoroughly mixed and diluted with mineral sediment contributed by solifluction (Gross, 1954, p. 198). Consequently there are few pollen diagrams recording the main part (Full-glacial, Pleni-glacial) of the last (Würm, Weichsel) cold phase. In Holland the first part of the Pleni-glacial is recorded by tundra pollen spectra, but the latter part is barren (Van der Hammen, 1952). For Denmark the mean July temperature for the Pleniglacial is estimated as 10°C below the present on the basis of the modern distribution of the plants represented

by pollen in the sediments (Andersen, DeVries, and Zagwijn, 1960). This phase was preceded in both Holland and Denmark by an Early-glacial phase with at least two interstadial intervals (Amersfoort and Brørup) during which subarctic and even north-temperate forests temporarily invaded (Andersen *et al.*, 1960); the pollen studies here are supplemented by radiocarbon datings that control the correlations.

For central and southern Europe the pollen investigations of the Pleni-glacial are less advanced, partly because of the lack of good pollen-iferous sediments dated by radiocarbon analysis. The Alpine glaciated region has yielded no good organic deposits for this phase. The best stratigraphic records of the last cold phase are found in the loess region of Austria and adjacent areas, but unfortunately neither the loess itself nor the intercalated soils contain pollen in sufficient quantities for analysis.

Paleozoological studies yield climatic information primarily through their implications about vegetation—forest forms *vs.* steppe forms *vs.* tundra forms. Animals are therefore one step further removed from a direct record of climate. Many of the important fossil mammal occurrences are at archeological sites (*e.g.*, caves), and there is the danger that some of the animals may have been imported by man from different vegetational zones and thus record hunting or dietary habits rather than the local climatic environment. Small mammals like rodents, which are ordinarily not dietary items, may be more significant in this respect. Other occurrences involve so few specimens that no reliable percentage analysis can be made. Nevertheless, mammal studies have their place in paleoclimatic analyses because, unlike pollen, the fossils may be found in loess, river terraces, and other sediments of glacial correlation.

Similarly the mollusk faunas yield information about vegetation and thus climate, and are particularly useful in the loess region where pollen analysis fails.

Some important mammal and molluscan faunas are enumerated in the section describing the climatic sequence for the late Pleistocene.

DISTRIBUTION OF PLEISTOCENE PERIGLACIAL FEATURES, AND RECONSTRUCTION OF PLEISTOCENE CLIMATIC ZONES

Reconstructions by Poser

Climatic zones. Although Soergel (1919) was probably the first to exploit *in toto* the possibilities of utilizing geomorphic and biologic data for reconstruction of European periglacial climate, the first detailed map and discussion of the distribution of Pleistocene frost soils and the reconstruction of climatic zones did not come until Poser (1948). Using what he considered to be three reliable indicators of permafrost, namely ice-wedge fillings, involutions, and asymmetric valleys, he mapped the distribution of Pleistocene permafrost and delineated several subzones within the entire periglacial area of central and western Europe.

The northern limit of Pleistocene permafrost was clearly the ice sheet. It has been mentioned that the ice apparently overrode already existing permafrost occasioned by the cooling climate, as indicated by the occurrence of involutions in sediments buried by young drift. But once this ground was insulated from the cold air by a thick ice cover, it is likely that it thawed by heat flow from the earth's interior, and when the ice withdrew from its maximum moraine the climate was not sufficiently severe to freeze the ground again, or in fact to cause enough solifluction to modify the deglaciated terrain very much. Consequently, the drift limit of the last glaciation in northern Germany and Denmark, as also in the Alpine foreland, is a striking geomorphic boundary between a relatively fresh glacial topography and a rolling topography on older drift molded by periglacial solifluction during the last glaciation (Gripp, 1932).

The southern boundary of permafrost during the last glaciation was likewise inferred by Poser (1948) from the limit of occurrence of the indicators used (Fig. 2). It extended from the Atlantic Ocean in Brittany eastward to the Seine Valley, thence south down the Rhone Valley and around the south base of the Alps. Throughout this course it followed just north of the northern forest limit that was inferred by Firbas (1939, Fig. 6) from the basal zone of Late-glacial pollen diagrams. The northerly position of both this forest line and the permafrost line in France is attributed to the ameliorating effect of the oceanic climate. [illegible]n as their modern counterparts are in Scandinavia and Alaska. The strong southward bend of the lines around the Alps reflects the combined cooling influences of the northern and Alpine ice masses. East of the Alps the permafrost line remained south of the Hungarian lowland because of the effect of the colder continental winters, but the forest line went north around the Carpathian Mountains because of the warm summers characteristic of the continental climate.

Three major climatic-vegetational zones are thus delineated by the relations of the southern limit of permafrost to the northern limit of the forest, as shown on Poser's map: (I) permafrost-tundra climate north of both lines, (II) continental permafrost-forest climate in the Hungarian lowland, and (IV) maritime forest climate without permafrost for the southern half of France. A fourth zone (III), a narrow strip across France, is problematical. Poser's data map (1948, Fig. 1) shows no basis for the demarcation of this zone. The pollen analyses used show treeless vegetation in Germany and Poland and birch-pine in Hungary; there is nothing indicated for France, so it is not clear why the forest limit was extended across northern France. More recent identification of the same permafrost indicators in France (Tricart, 1956a) places the permafrost boundary actually in the south of France rather than in the north, so the concept of a broad zone of maritime forest climate without permafrost over all but southernmost France is not justified if these indicators are accepted.

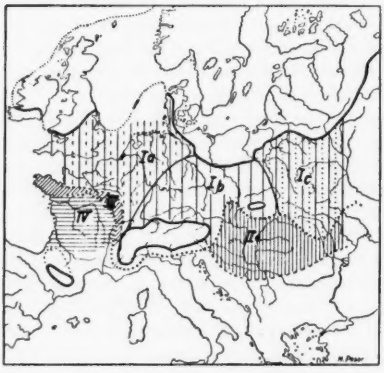

Figure 2. Climatic provinces of central and western Europe during the Würm glacial phase, according to Poser (1948, Fig. 6). The heavy solid line and its dotted continuation delimits the ice sheets, the dashed line shows the southern limit of permafrost, and the line of circles indicates the northern forest border. The following provinces are identified:

I Permafrost-tundra climate
  Ia Glacial-maritime province
  Ib Province between the ice sheets
  Ic Glacial-continental province
II Continental permafrost-forest climate
III Maritime tundra climate without permafrost
IV Maritime forest climate without permafrost

The use of the basal zones of pollen diagrams to characterize the vegetation at the time of the maximum of the last glacial period requires many assumptions and involves a consideration of the time factor, which seems to have been neglected in most of the climatic reconstructions based on frost features. The Alleröd zone is the oldest zone that can be identified in Late-glacial pollen diagrams in most of the parts of Europe that were not covered by ice during the last glaciation, and it is claimed to be identified over hundreds of localities (Gross, 1954, Fig. 1; Wright, 1957, p. 449 ff.). Although there is always the problem of the proper correlation of the substratum beneath a pollen-bearing deposit, especially in unglaciated regions, one is impressed by the similarities among the very large number of available detailed diagrams for localities distributed across several vegetational provinces from the young glaciated region in the north (where the substratum is known) to the Alps and southeastern Europe (Firbas, 1949, p. 301). The fact remains, however, that the pollen zone below the Alleröd usually represents an unknown time interval prior to about 12,000 years ago. It may not reflect the vegetational conditions at the time of maximum cold of the last glacial period, but rather some time later when the climate had already improved considerably. The 16 localities spotted on Poser's map in central and western Europe presumably represent those in which the basal pollen zones show treeless vegetation. If the region was treeless at this time during the warming stages near the end of the last cold phase, it must have been treeless during the maximum cold, so Poser is on safe ground as far as the mapped treeless area is concerned. But for the areas mapped as wooded, synchroneity is not assured unless independent dating is possible.

Furthermore, it is not known when the frost features were formed within the last cold period, which started perhaps 70,000 years ago and was marked by climatic oscillations of varying magnitudes (see p. 959). In the loess region of Austria and adjacent areas a phase of solifluction preceded each phase of loess deposition, and small ice-wedge fillings and involutions have been found in the solifluction loess of the

more humid parts of southern Czechoslovakia, but not the drier (Brandtner, 1956, p. 132). Adequate moisture as well as cold was apparently necessary for the development of frost features. The paleontological studies and the correlation of loesses with glacial deposits indicate that the coldest time came with the maximum of glaciation near the end of the last cold phase, whereas the solifluction loess was formed during the early part. These relations imply that frost features may have formed throughout the cold period, but that during the time of maximum cold they were restricted to relatively moist sites.

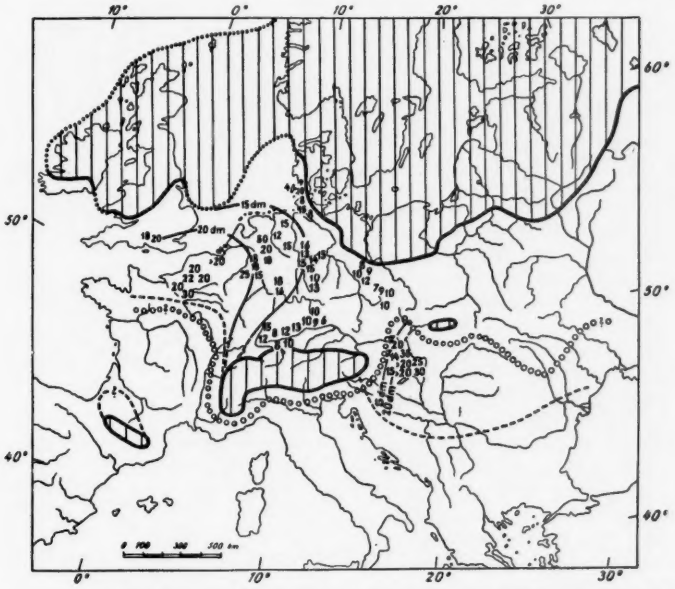

Figure 3. Depth of summer thaw in central Europe during the Würm glacial phase as based on the depth of involutions (Poser, 1948, Fig. 2). Figures in decimeters. Ruled area shows extent of ice sheets. Dashed line indicates southern limit of permafrost, and line of circles shows northern limit of forest.

In order to characterize further the climate of Europe during the last glaciation, Poser assumed that the southern limit of permafrost at that time corresponded to the −2°C annual isotherm, and that the northern forest border fell along the 10° July isotherm. Where the two lines cross at the east end of the Alps, the entire annual temperature cycle can be calculated from these two control temperatures. Thus if the July mean was 10°C and the annual mean −2°C, the January mean must have been −13°C. For northern France, where the two Würm lines are supposedly close together, Poser made similar calculations to show that the seasonal contrast was not so great (just as today) because of the maritime influence. He also attempted to show that in France the summer temperatures were lowered in the Pleistocene (with respect to today) more than the winter temperatures, whereas in Hungary, which has a continental climate, the winter depression was much greater than the summer depression. Also, the Pleistocene depression of the annual mean temperature was much greater in Hungary (by 10°C) than in France (by 4°C) because of this same effect. Although the new data on the position of the Würm permafrost line in France (Tricart, 1956a) will require recalculation of the temperatures, the contrasts between the maritime and continental climates will probably still be noticeable.

Summer temperature. Poser (1948) also plotted from the literature the depths at which involutions have been found at 71 localities

throughout Europe (Fig. 3). The map is intended to show the range of thaw depth and thus the summer temperature, on the theory that involutions are formed near the base of the active layer of permafrost. Shallow thaw depths, less than 15 decimeters, occur in southern England and in Denmark and northwest Germany because of the cooling effect of the ice sheet in summer. Shallow thaw depths are also found through central Germany to the Alps, presumably because of the combined cooling effects of the northern and Alpine ice sheets, which actually were no more than 300 miles apart at their closest points. Cold summers in this region are further implied by the absence of a forest cover, as indicated by pollen analyses.

Thaw depths greater than 20 dcm are inferred farther west in northern France. Here the maritime influence brought summers that were longer (although perhaps cooler) than the more continental regions to the east. Greater thaw depths were also found in the Hungarian lowland. Here the location deep within the continent brought strongly contrasting seasons. Winters were cold enough for permafrost even at this low latitude, but the correspondingly warmer summers resulted in relatively deep summer thaw as well as the development of forest—the one area in Europe characterized by both permafrost and forest, according to Poser.

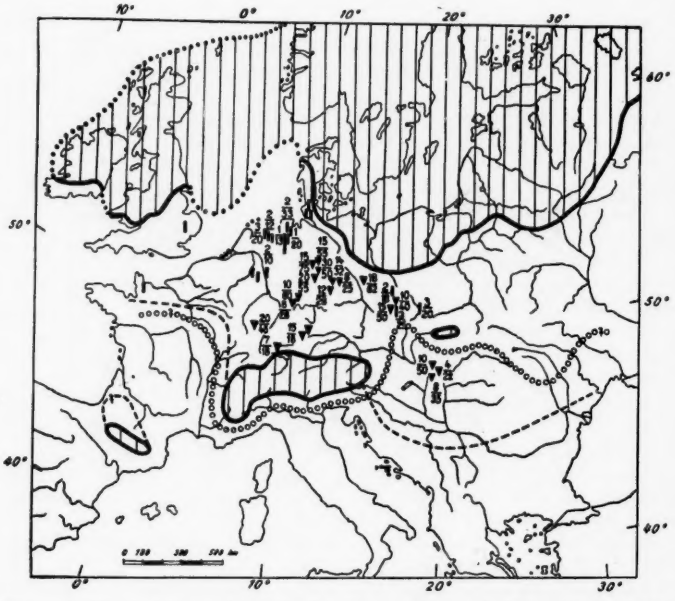

Figure 4. Winter frost cracking of soil in Europe during the Würm (Poser, 1948, Fig. 4). Ruled area represents ice sheets. Dashed line shows southern limit of permafrost, and line of circles shows northern limit of forest. Small triangles show occurrences of loam wedges, and small black rectangles indicate frost cracks. Pairs of numbers show the breadth (above) and depth (below) of wedges and cracks, in decimeters.

Winter temperature. Another map by Poser shows the distribution of about 25 localities of ice-wedge casts that have been reported in the literature for Europe (Fig. 4). The depth and the width (at the top) are noted for each locality and are believed to be directly proportional to the severity of the winter cold. Although the observations are not sufficiently numerous to justify lines of equal wedge depth and width, Poser infers several regions of differing winter climate.

Close to the North Sea the wedges are less

than 30 dcm deep and only a few decimeters wide at the top. Here the maritime influence kept the winters relatively mild and short, despite the fact that the ice sheet was not far to the north. Thus the penetration of winter cold and the consequent ground contraction was not great. The total depth of permafrost itself is unknown; it is determined by summer as well as winter temperature.

In central Europe the depths of the wedges

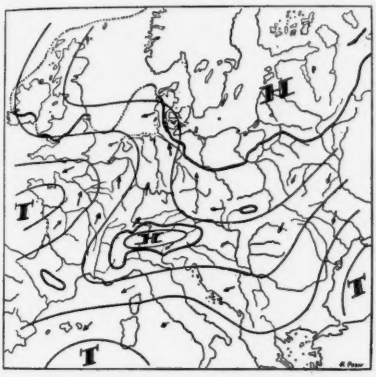

Figure 5. Summer atmospheric pressure and wind directions in Europe during the Würm glacial phase, as determined from the climatic and vegetational provinces (Poser, 1948, Fig. 3). H = high-pressure area, T = low-pressure area

are generally 50–80 dcm, and the widths are up to 30 dcm. Such large features are believed to reflect severe winter cold with deep and frequent cracking of the ground. The lower winter temperatures were caused by the junction of cold air masses over the north European and Alpine ice sheets, intensifying the increasing continentality of the interior region.

In the Hungarian lowland a third climatic region is identified, although on the basis of only three observations. Here the wedges are deep (22, 35, 50 dcm) but relatively narrow (4, 8, 10 dcm respectively). Severe winter cold is consistent in a continental climate, but infrequent cracking may reflect the insulation provided by a forest cover.

Atmospheric pressure. On the basis of the location of the three major Pleistocene climatic zones thus delineated in Europe—permafrost-tundra zone across the north, permafrost-forest zone around the Hungarian lowland, and forest zone without permafrost in France—Poser (1948) mapped the summer and winter atmospheric pressure patterns and wind directions for the Pleistocene (Fig. 5). A high-pressure area is located over the Scandinavian ice sheet, with a ridge connection across Germany to another high over the Alps, especially in winter. A low-pressure cell is postulated for the warm Hungarian lowland in the summer, but displaced eastward to the Black Sea region in winter by the expansion of the Asiatic continental high. This low would have pulled dry air from the northeast and east off the continent—incompatible with the presence of forest in the Hungarian lowland. Perhaps, however, the same low-pressure cell attracted cyclonic storms which must have entered the Mediterranean from the west much more frequently in the Pleistocene than today, and moisture could have been brought to the Hungarian lowland by this route. On the other hand, vegetational reconstructions by Büdel (1951) and Frenzel and Troll (1952) show the Hungarian lowland as nonforested, so perhaps there is no problem to explain here.

Other low-pressure cells are postulated over the western Mediterranean and over France. The former would have attracted Atlantic storms along the Mediterranean, and the French low would have allowed moist air and storminess to invade northwestern Europe for ice-sheet nourishment in both winter and summer. The moisture must not have penetrated far eastward into central Europe, however, for the climate was distinctly dry in lower Austria and adjacent areas during the principal times of loess deposition, although important regional fluctuations of precipitation are suggested by the humid-type solifluction phase at the base of each loess bed.

Wind and pressure systems. A further analysis of Pleistocene atmospheric pressure patterns and wind directions was made by Poser (1950) on the basis of the study of Late-glacial sand dunes in northern and eastern Europe. Although this study does not give a picture of the wind systems at the time of maximum frost activity, maximum loess deposition, or even maximum glaciation, it reveals the situation when the Scandinavian ice sheet was still in northern Germany or in southern Scandinavia, prior to reforestation.

Wind directions at the time of dune formation are inferred by Poser from dune form and internal structure. It is important whether the dune-forming winds represent prevailing sum-

mer winds or merely storm winds caused by passing cyclones. In the former case the pattern of dunes and thus prevailing winds might reveal the mean summer pressure pattern for Late-glacial time. Poser determined that the sand of the mean size grade concerned could all be moved with a wind strength of 8–12 miles per hour for variable wind, which prevents packing of the surface layer of sand grains, or 25–31 miles per hour in the case of constant winds and a packed surface. The presence of a few coarse laminae in the dunes indicates that strong winds did blow from time to time, and brought coarse sand from available outwash plains or other sources. The dominance of smaller grains and the uniformity of particle size, however, suggest that the dune-forming winds were relatively constant, and a threshold of 19–24 miles per hour is taken for the velocity required for transport.

The dunes in northeastern Germany and Poland are primarily bow-shaped dunes; those of Hungary are linear dunes. The uniformity and simplicity of form and structure suggest to Poser that the winds were not the changeable winds associated with passing cyclonic depressions, but rather a wind system around an anticyclone. A further indication of this relation comes from the fact that modern dune-forming winds in northern and eastern Europe have precisely the same mean directions as the prevailing wind directions inferred for the Late-glacial dunes, namely southwesterly winds in Belgium, westerly in eastern Germany, northwesterly in Poland, and northerly in Hungary. Examination of modern weather maps showed that on most of the days during which the wind blew 19–24 miles per hour at one or another of the regions under consideration the map showed a high-pressure area over Europe, as an eastward extension from the Azores High. In fact the most favorable circumstances seemed to involve some storminess around the high, to produce moderately variable winds more conducive to sand transport.

The pressure map (Fig. 6) reconstructed by Poser for the Late-glacial summer resembles the summer map he had drawn for the Würm maximum on the basis of the distribution of certain frost soils (Fig. 5), except that in the Late-glacial the Scandinavian high-pressure area has retreated to the north along with the ice sheet that produced it, and is no longer connected with the high over central Europe. Compared to the summer pressure map of today, the high over Europe is considered to be more stable because of local cooling influence by the ice sheet and by cool glacial meltwater in the sea off the coast of western Europe. That the region was still cool cannot be doubted from the occurrence of tundra plants in the Late-glacial pollen zones. The extent to which depressions moved along the North Sea and Baltic Sea at this time cannot be easily determined, but the persistence of a European high into the Late-glacial implies that some storms still took the Mediterranean route south of this air mass.

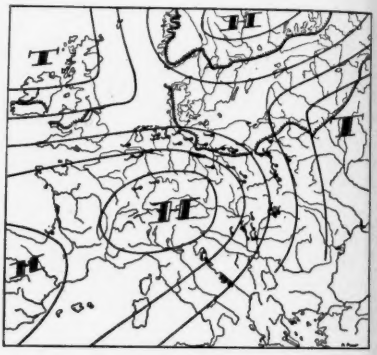

Figure 6. Summer atmospheric pressure and winds during the Late-glacial phase of the Würm, based on morphology of ancient sand dunes (Poser, 1950, Fig. 1). Hachured lines show the positions of the north European ice sheet at the times of the beginning and the end of the Late-glacial.

Reconstructions by Büdel

Snow line and tree line. Climatic zones of Europe for the last cold period were reconstructed with a different basis and with more detail by Büdel (1951). He assumed that the polar forest border in the Pleistocene followed the 10.5°C July isotherm just as it does today and that this isotherm in turn followed the 1000-m level of snow line in the Pleistocene just as it does today in the polar regions. The level of Pleistocene snow line was determined in the mountains of Europe by Brusch and Büdel on the basis of published field data (Büdel, 1949, Fig. 3). The 1000-m Pleistocene snow line thus reconstructed extends across the Central Plateau of France and across the German Mittelgebirge. The position of the

Würm 10.5°C July isotherm in the nonmountainous areas was then determined by applying the topographic correction factor (lapse rate) of 0.5°C/100 m of elevation. The corrected line extends from the mouth of the Garonne River in France southeast to the French Riviera, protrudes up the Rhone Valley, and then extends to the south base of the Alps. The Pyrenees are surrounded by a separate line (Büdel, 1951, Fig. 2).

Having made the assumption that the Pleistocene polar forest line followed the 10.5°C July isotherm and the 1000-m snow line, Büdel then points out that Pleistocene tree-line depression in southern Europe in fact exceeded the snow-line depression (Firbas, 1939, p. 105), and he thereupon proceeds to make exceptions to his assumption. Thus he draws the polar forest limit, as determined from "paleobotanical investigations," south of the Pyrenees and other mountains of the Iberian Peninsula, thence along the French Riviera to the Alps. This line thus essentially excludes the Garonne-Aquitanian Basin and the lower Rhone from the Würm forested region despite the fact that these areas were south of the reconstructed 10.5°C July isotherm. A later map by Büdel, here reproduced as Figure 7, contains slight revisions in southern France, with the omission of the line for the 10.5°C isotherm.

Inasmuch as the polar forest limit in the west was placed by Poser in northern France but by Büdel in central Spain, 400 miles away, both on the basis of "paleobotanical investigations," one wonders how accurately the Pleistocene forest limit can be determined. Neither author shows the distribution of the paleobotanical sites which guided the mapping. Firbas's (1939, Fig. 6) boundaries of Würm forest zones do not extend westward from Germany and Italy. Gross's (1954, Fig. 1) map of Alleröd localities shows very few in France, and if the Alleröd cannot be identified in a pollen diagram there is generally no datable base, at least in diagrams for sites in unglaciated regions. It appears that Büdel shifted the forest limit to the south because of Cailleux's discovery of frost features, including ice-wedge fillings and involutions, in central and even southern France, in the belief that these features could not have formed under continuous forest cover. This is not paleobotanical evidence for a forest limit, but geomorphic. Actually, as Hopkins *et al.* (1955) have shown in Alaska, such frost features can be found today in regions of sporadic and discontinuous permafrost, which are also the regions in which the forest gives way to tundra. Further, the forest limit is not a line but a zone. Büdel offers a compromise about France by inserting a transitional zone of forest tundra or shrub tundra north of the "forest limit." This zone would presumably resemble the park tundra of the pollen analysts. On Büdel's map it includes the southwestern quarter of France and all of the Iberian Peninsula north of the forest limit (except the high Pyrenees).

One of Poser's major climatic provinces was based on a forest-covered France. Poser advocated strong climatic influence exerted by the two ice sheets and attributed the forest zone on the west to the ameliorating effect of the maritime climate. Büdel, on the other hand, uses the absence of forest in France to support his thesis that the Pleistocene climate of Europe was little affected by the ice sheets, pointing out that the tree line in western Europe was 700 miles from the ice sheet, but in Russia was less than 100 miles away. Studies of the form, distribution, and other characteristics of sand dunes and loess deposits all indicate that the dune-forming winds in western Europe came from the ocean on the west, not from the ice sheet, which therefore could not have influenced the climate appreciably except in the frost-rubble tundra zone immediately bordering the ice. This is a forceful argument, and further field evidence on the Würm climate and vegetation of France will be welcome. The recent map of Würm frost features by Tricart (1956a, Pl. 3) is impressive in its detail; one cannot be too skeptical about climatic interpretation of stabilized *eboulis* (talus and slope rubble) when there are so many occurrences of similar concentrated rock debris in demonstrably Pleistocene archeological levels in caves, and when there are so many occurrences of ice-wedge fillings and involutions in the same area (unless one is skeptical about the significance of these features as well). One must accept the French map on its face value, for there is no way to evaluate the correctness of the interpretation of individual exposures. The French workers are enthusiasts for the periglacial because of their experience with frost features in Greenland and Iceland. One comes away with the conclusion that France may have been quite cold during certain parts of the Würm, with sporadic permafrost possibly extending far to the south. The faunal evidence from some of the Upper Paleolithic sites in southern France certainly suggests severe climate. We must await further pollen studies,

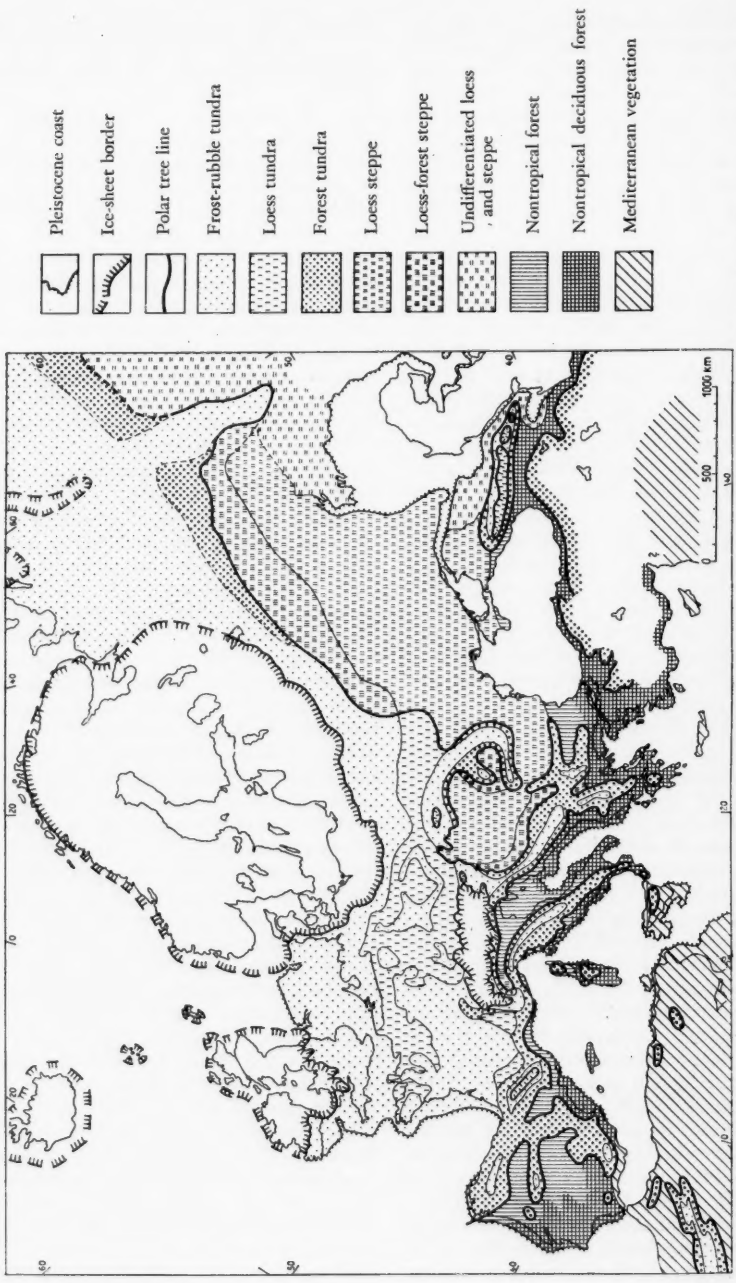

Figure 7. Vegetational zones of Europe during the Würm glacial phase, according to Büdel (Woldstedt, 1958, Fig. 129, slightly modified by Büdel from his 1951 map)

however, before drawing the position of the polar forest limit or tree line with more certainty.

Eastward from France, Büdel draws the polar forest line as coincident with the 10.5°C July isotherm. The line skirts the Apennines, Dinaric-Balkan Range, and the Carpathian arc and then extends northeastward into Russia. Its position in the mountains of southern and southeastern Europe was delineated by topography rather than by field evidence.

Another possible discrepancy in the location of the Pleistocene forest line comes in the Hungarian lowland. Poser mapped the region as forested and showed six localities in which birch-pine characterizes presumably the lowest parts of the pollen diagrams. Büdel, on the other hand, maps the area as steppe, referring only to paleobotanical investigations in support.

Climatic zones. The polar forest line is the key to the climatic zonation of Büdel. In western and central Europe he recognizes a transition zone on the north (shrub tundra), but to the east across most of Europe the tundra zone is separated into loess tundra and frost-rubble tundra. The frost-rubble tundra zone occurs as a continuous belt close to the ice sheet from Ireland to Russia, and also occurs on the upper parts of the highlands of western and central Europe. In this zone some loess may have been deflated, for certainly winds and suitable source areas were present, but continual frost disturbance of the soil did not permit its accumulation as pure loess, according to Büdel.

The loess belt occurs at lower elevations through western and central Europe. Its upper limit climbs from 1000 feet above sea level in the north-northwest to 2000 feet at the north base of the Alps, rising just like any other boundary controlled by vegetation. The vegetation is considered to be tundra, with a mat thick enough to inhibit continual frost disturbance of the loess that was deposited. A transitional zone occurs near the upper boundary in the mountains, where loess and solifluction deposits are interbedded. Fluctuations of the geographic boundary may also be inferred from loess sections, in which each loess cycle, at least in the more humid parts of Europe, commonly starts with a solifluction loess, grades into pure loess, and terminates with a soil. Thus the frost-rubble tundra zone with a cooler, more humid climate moved down even into the lowlands at the beginning of a cold phase.

The loess belt as a whole broadens from Brittany on the west to the Ukraine on the east, with interruptions caused by highlands. As far east as the Russian border it is placed by Büdel in a loess tundra zone, but here northeast of the Carpathian Mountains it is crossed obliquely by the Pleistocene polar forest line. The forest line in this continental region, however, was probably a rather indefinite feature. Certainly a distinction between loess tundra west of this line and loess steppe (with a few trees) east of this line must be tenuous.

Büdel distinguishes between tree line and forest line. These two boundaries were considered continuous across southern Europe to the Carpathians, but in the east, where the dry continental climate prevailed, the continuous forest was confined to the south, in fact south of the Black Sea. Northward was the broad loess steppe, which had an increasing number of trees northward (less evaporation) until a narrow loess-forest steppe zone was reached just south of the tree line. The tree line here in Russia was scarcely 100 miles from the ice sheet, presumably because of the warmth of the continental summer and despite the proximity of the ice sheet, which Büdel, of course, believes had relatively little effect on the "periglacial" climate.

Thus in Russia the climatic-vegetation zones and boundaries from north to south were frost-rubble tundra, shrub tundra, tree line, loess-forest steppe, loess steppe, forest line, and forest. In western Europe they were frost-rubble tundra, loess tundra, tree line, shrub tundra, forest line, and forest.

The Würm vegetation zones in European Russia were sketched largely in speculation by Büdel, but were drawn with more refinement by Frenzel and Troll (1952) on the basis of more extensive Russian pollen-analytical, plant-geographical, and geomorphic studies and were extended far eastward to incorporate the earlier work of von Wissman (1938) for China. On this map the vegetation zones for central and western Europe were prepared in consultation with Büdel and differ little from those of the earlier Büdel (1951) map.

In eastern Europe, however, the following zones were differentiated by Frenzel and Troll (1952). Close to the ice was a narrow frost-rubble tundra zone with dwarf birch and willow, increasing southward. Then came the "tree line," south of which was a mixed zone of forest-tundra herbs but also chenopods, *Artemisia*, and other steppe elements. Büdel had termed this zone simply a forest steppe and considered it much narrower. Ice wedges are

described from this zone, as well as solifluction features, and the southern limit of sporadic permafrost is put at a line extending through Stalingrad.

South of the mixed tundra-forest-steppe zone is the steppe proper (loess steppe of Büdel), with pine and spruce on favorable sites. In the northern part of the zone there are believed to have been outliers of birch and willow shrubs and, in the south, long extensions of deciduous bushes up the river lowlands.

The vegetational zonation in Russia during the last glaciation does not differ from that of today to the extent that it does farther west because the climate has been continental throughout. During the glacial period the winters were colder, as indicated by the southward extension of permafrost. Also, the summers were not so warm, and thus tree line was also pushed somewhat to the south. But the summer change in particular was probably not so pronounced as it was in central Europe, where the combined effect of the Scandinavian and Alpine ice sheets intensified the refrigeration.

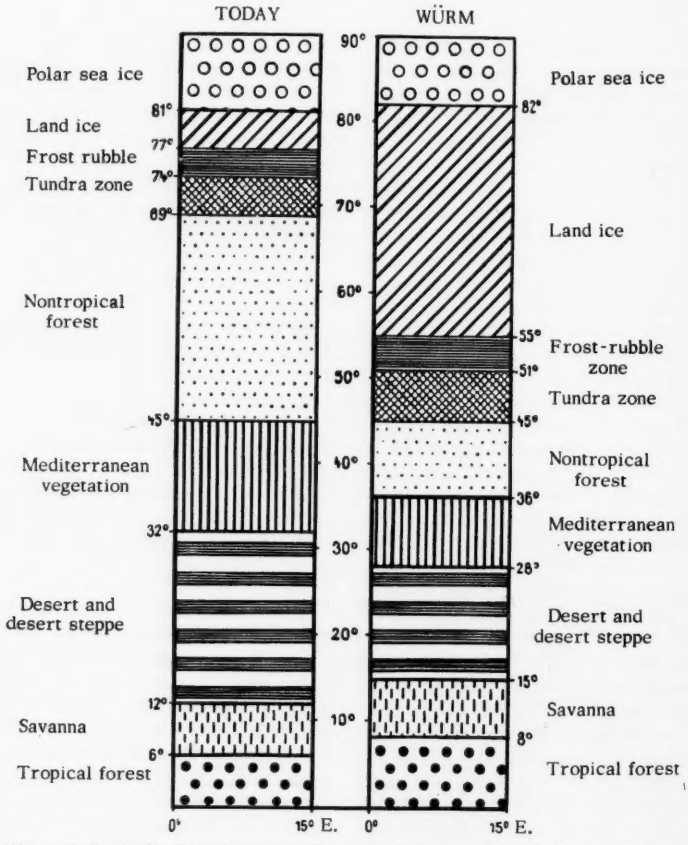

Figure 8. Latitudinal displacement of terrestrial climatic zones during the Würm in the meridional belt 0°–15° East Longitude (Büdel, 1951, Fig. 1).

The equatorward displacement of the vegetational zones of western Europe during the Würm are represented diagrammatically by Büdel (Fig. 8). The most pronounced changes were in the zone of land ice, which was in-

creased from a breadth of 4° to 27° of latitude, and the zone of nontropical forest, which decreased from a breadth of 24° of latitude today to 10° in the Pleistocene. The zone of Mediterranean vegetation was pushed to the south and was narrowed, and the subtropical desert and desert steppe were decreased in breadth from 20° of latitude to 13° not only because of restriction from the north but also because of the expansion of the zones of tropical forest and savanna on the south, according to Büdel. The extent of the changes postulated for lower latitudes are based on Büdel's work (1952) in northern Africa and are beyond the area under consideration.

SEQUENCE OF CLIMATIC CHANGE

Introduction

In the last decade there has been renewed interest in problems of the classification of the last (Würm, Weichsel) cold phase of the Pleistocene and in the correlation of sequences among the three principal areas concerned—the north European glaciated area, the Alpine area, and the periglacial area between. The interest has been impelled by new field studies particularly in the loess region, and by the application of the radiocarbon dating technique to carefully collected samples from sections that are significant stratigraphically or geomorphically. This gives the possibility for the first time of correlation on an absolute time scale. Particularly important here are the analyses from the Groningen radiocarbon laboratory in the time range 30,000–65,000 years ago, made possible by isotopic enrichment of samples before counting, for these analyses bring within dating range the materials from the early part of the last cold phase of the Pleistocene. The new work, however, has not yet settled a long-standing controversy about the basic classification of the last cold phase, whether it be single or multiple, and in fact the controversy seems to have become more vigorous than ever. Not until this question is settled will a general climatic curve for the late Pleistocene be accepted.

Penck Classification

In the Alps, Penck and Brückner (1909) mapped the Würm moraines generally as three outer loops at the base of the mountains plus some smaller moraines far up the valleys. Although they recognized possible interstadial fluctuations (Laufen and Achen) in the Würm prior to the construction of the main moraines, Penck (1922, p. 240) later reassigned these to the preceding interglacial, and thereafter considered the Würm to represent essentially a single climatic cycle, with minor fluctuations during ice retreat. The Würm outside the moraines is represented by the Niederterrasse in the Alpine valleys that lead to the Rhine and Danube rivers. In northern Europe the equivalent Weichsel moraines include several ridges across northern Germany and southern Scandinavia, *e.g.* Brandenburg, Frankfort, Pomeranian, Langeland, Salpausselkä moraines. The later fluctuations in the sequence are correlated with the three tundra ("Dryas") phases bounding the Bölling and Alleröd oscillations recorded in Late-glacial pollen diagrams (Wright, 1957).

Penck's single-cycle classification of the Würm has been defended repeatedly by Büdel (1953, p. 264), who believes further that superimposed on the single temperature cycle there was a precipitation change—from humid in the first part of the cold phase to dry in the second—as indicated especially by the solifluction loess and the overlying pure loess of the periglacial area (Fig. 9). Other champions of this basic classification are Weidenbach (1953) and Graul (1952), who have mapped the moraines and terraces of the Alpine foreland, and Freising (1957), Brunnacker (1956), and Fink (1959), who have studied the loess stratigraphy of southern Germany and Austria.

Soergel Classification

A different scheme of subdivision of the last major cold phase came into use following the studies of Soergel (1919) of the loesses, buried soils, terraces, frost soils, and mammal faunas of central Europe. The classification was based primarily on the loess stratigraphy, in which the so-called Older Loess was correlated with Riss glaciation of Penck and the Younger Loess with the Würm glaciation. The Younger Loess was subdivided into Y.L. I and Y.L. II by a buried soil, and Y.L. II into IIa and IIb in turn by a less well developed soil. The terms Würm I, II, and III were soon applied to the three units of Younger Loess in the belief that Y.L. II (Würm II) correlated with the main outer young moraine loops of the Alpine glaciers. In this scheme Würm I was represented in the Alps by a moraine in front of the main Würm II moraines but inside the Riss moraines (Soergel, 1919, p. 89), and in the north the Würm I was represented by the Warthe drift in front of the Brandenburg moraine. Weidenbach (1937, p. 87), however, has shown that the so-called

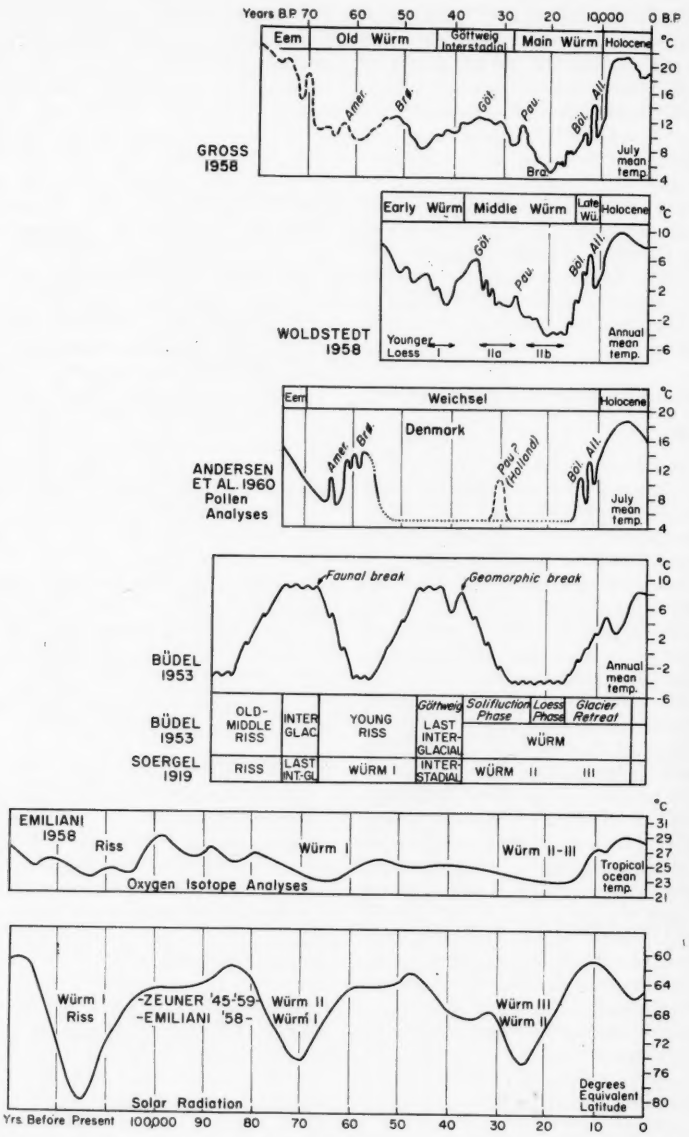

Figure 9. Climatic curves of the late Pleistocene according to various authors. Amer. = Amersfoort, Brø. = Brørup, Göt. = Göttweig, Pau. = Paudorf, Bra. = Brandenburg, Böl. = Bölling, All. = Alleröd

Würm I moraine of Soergel relates to the Riss complex and the Riss terrace (Hochterrasse) rather than to the Würm, so should be termed Young Riss rather than Würm I, and in the north Woldstedt (1958, p. 20) demonstrated that the Warthe belongs to the Saale glaciation rather than to the Weichsel. These denials of a Würm I moraine in front of the Würm II moraines left the Würm I without a representation in the glaciated areas, so an alternative proposal was made that the innermost of the three Würm moraine loops in several Alpine valleys is actually a Würm I moraine overridden during later readvance of the ice in the Würm II stadial (Eberl, 1930; Weinberger, 1955, p. 15). The interstadial significance of this feature is problematical, however, because of the lack of any weathering horizon or other evidence of appreciable ice retreat between the time of formation of the "overridden" moraine and the main Würm outer moraines.

The Soergel classification, that the Würm is bipartite or tripartite, was adopted by Zeuner (1945–1959), who related the inferred Würm temperature curve to the Milankovitch radiation curve for the last 100,000 years. It has been utilized almost exclusively by archeologists in the interpretation of cave deposits. It recently has been greatly elaborated by detailed studies in the loess region of Czechoslovakia (Prosek and Lozek, 1957) and Austria (Brandtner, 1956), and has been adopted in the voluminous reviews of Gross (1956; 1958; 1959) and in the authoritative book of Woldstedt (1958, p. 243–248) as the key to the climatic sequence of the late Pleistocene. The matter is by no means settled, however, despite the vituperation of Gross (*see* comments to Movius, 1960, p. 378) and warrants review herewith because it concerns the form of the climatic curve for the late Pleistocene.

Climatic Curve of Gross

General. In his attempt to summarize, generalize, and correlate, Gross adapted the terms Krems, Göttweig, and Paudorf that had been used by Götzinger (1936) with different correlation for three soils in the loess sequence in a small area about 30 miles west of Vienna close to the Danube River. Whereas Götzinger had used the correlations Mindel/Riss, Riss/Würm, and Würm I/II respectively, Gross (1956), basically following Brandtner (1956), moved the sequence up one step to Riss/Würm, Würm I/II, and Würm II/III by assigning the soils to the weathering intervals Older Loess/Younger Loess, Y.L. I/II, Y.L. II a/b respectively in the Soergel classification. The three terms Krems, Göttweig, and Paudorf have since been widely adopted (Woldstedt, 1958, Fig. 86; Movius, 1960) and assigned the following dates on the radiocarbon time scale: Krems (Eem) >65,000 years ago, Göttweig 44,000–28,000, and Paudorf 25,000. The climatic curve of Gross (Fig. 9) and the similar curve of Woldstedt (Fig. 9) for the entire Würm, as based on these subdivisions and this C^{14} time scale, assumes temperature values derived in part from molluscan and mammalian faunal studies largely summarized from the Czech loess studies by Brandtner (1956, p. 155).

Loess stratigraphy. The loess stratigraphy of four general areas in western and central Europe has been studied intensively in recent years: (1) northern France (Bordes, 1954) and Belgium (Tavernier and deHeinzelin, 1957); (2) southern Germany (Schönhals, 1951; Freising, 1957; Brunnacker, 1956); (3) Lower Austria (Brandtner, 1954; 1956; Fink, 1956; 1959); and (4) Czechoslovakia (Prosek and Lozek, 1957). Correlations have been attempted between France and Germany (Bordes and Müller-Beck, 1956), France and Czechoslovakia (Valoch and Bordes, 1957), and Austria and Czechoslovakia (Brandtner, 1956). The following review is confined largely to the loess stratigraphy of Czechoslovakia and Lower Austria because the sequence is more completely developed, radiocarbon dating and molluscan studies have been applied, and part of the area shows connection with the Alpine glaciofluvial features.

The Czech loesses are related to periglacial terraces of river systems draining to the area of north European glaciation via the Elbe River (Prosek and Lozek, 1957, p. 43–54), and the correlation of loesses and soils depends to a great extent upon the proper correlation of these terraces with glacial events in the north. Connections southward with the Danube terraces are unfortunately either not well developed or are controversial (Brandtner, 1956, p. 166–172).

On the Moldau River near Prague the terrace and loess that are correlated as Saale (Riss) bear a soil (termed Riss/Würm and correlated with Krems) with which is associated a snail fauna characterized by *Helicigona banatica* and other forms now found in warmer regions—the so-called "Banatica" fauna, believed to be the last true interglacial molluscan fauna of the loess region (Prosek and Lozek, 1957, p. 74–

86). The overlying Younger Loess I contains *Helicella striata* and other forms that indicate a relatively warm steppe environment ("striata" fauna).

The soil found on this younger Loess I is termed Würm I/II and correlated with the Göttweig soil of Austria. It is considered to be as well developed as the "Krems" soil below, according to Prosek and Lozek (1957, p. 77), but not so according to Brandtner (1956, p. 136), who considers that the Czech workers have misinterpreted the soil profiles. At any rate, the soil differs from the "Krems" to the important extent that the associated snail fauna is not interglacial like the "Banatica" type but is essentially the "Striata" fauna as in the underlying Younger Loess I. Also present, however, are a few forms suggesting open deciduous woodland and thus somewhat moister conditions, with temperature much like the present (Prosek and Lozek, 1957, p. 77). Inferences from snail faunas concerning the climate at the time of soil formation must be limited because of the fact that few fossils are preserved in the soil itself, especially under humid conditions.

Younger Loess IIa contains the "Striata" fauna at the base; this is replaced upward abruptly by the "Columella" fauna, characterized by *Columella edentula columella* and other arctic-alpine forms, indicating a change to very cold steppe or tundra conditions. This fauna persists through the slight interruption of the Würm II/III ("Paudorf") interstadial and culminates in Younger Loess IIb.

The succession of mammal faunas in the loess deposits is incomplete. Most of the material is from archeological sites, many of which cannot be related to units of the loess stratigraphy. The information is supplemented by evidence from cave sites, but here the correlation must be based on independent methods, such as the physical stratigraphy of cave deposits—frost-rubble layers, weathering horizons, carbonate zones, analyses of particle size and shape, and certain other features often of problematical significance. Correlation is also made (precariously) by means of the cultural stages themselves. The possible errors that creep into mammalian faunal lists for the various units of the late Pleistocene are therefore many. The extended table produced by Mottl (1953; *see also* Flint, 1957, p. 451–455), for example, is difficult to evaluate.

A generalized succession of mammal faunas applicable for the Austrian area (Brandtner, 1956, p. 164–165) shows for the Early Würm a temperate steppe fauna of mammoth (*Elephas primigenius*), woolly rhinoceros, giant deer (the so-called "Primigenius" fauna), along with hyena, bear, cat, wild horse, and bison. This fauna thus parallels the "Striata" snail fauna. The Göttweig soil in the loess (and in cave layers correlated therewith) yields remains of red deer, moose, red fox, brown bear, badger, marten, and lynx, which imply at least some woods on favorable sites. Ibex, chamois, polecat, ermine, weasel, and ground squirrel, although associated with culture layers and thus potentially imported, imply at least temperate climatic conditions. The presence as well of steppe elements of the "Primigenius" fauna suggests an overall park landscape for this interval. Above the Paudorf soil, along with the cold "Columella" snail fauna, such tundra forms as reindeer, lemming, arctic fox, snow hare, white grouse, wolf, and wolverine appear, and the steppe animals are absent.

Pollen analyses, which generally give the most direct and detailed data about contemporary vegetation, have provided little information about the relations in the loess region because few satisfactory sites with well-preserved pollen have been found. Insufficient quantities of pollen are present in the loess itself, and in the intervening soils the pollen is poorly preserved. Schütrumpf (Brandtner, 1956, p. 150) reports an analysis from the humic bands over the "Göttweig" soil at Unterwisternitz in southern Czechoslovakia as containing mostly pine and nonarboreal elements. A buried peat in loess in the area of the Saale (Riss) moraine at Cimoskowicze in central Poland is considered by Woldstedt (1958, p. 130) as the best case for a complete "Göttweig" pollen profile. The loess here rests on a lower peat that shows an Eem-type (last interglacial) pollen diagram, so the upper peat, which reveals vegetation of dominant pine, is correlated as Würm interstadial (*i.e.*, Göttweig).

Loesses and terraces of the Alpine foreland. The critical area for the Würm classification is the Alpine foreland, where the glaciofluvial terraces are related to the loess stratigraphy. In southern Germany and western Austria the Alpine glaciers came down to the Rhine and Danube River valleys, and the moraines and terraces are well developed. The loesses in this region are thin, however, and the soils have been deeply truncated or transported by processes of solifluction during the beginning of the cold phase following their formation, for

the climate here in the west was relatively humid, as it is today. The modern soil in this region is a Parabraunerde, similar to the Gray-Brown Podzolic soil of American terminology. The annual precipitation is now 30–35 inches (Fink, 1956, p. 52). Few stratigraphic sections contain more than one buried soil. Although the loess stratigraphy has been carefully studied in the upland area north of the Danube River (Freising, 1957; Brunnacker, 1956; Schönhals, 1951), correlations with the Alpine terrace sequence and with the Austrian loess soils are speculative.

Farther east the Alpine glaciers did not reach the foreland, but the main terraces may be traced along the Danube as far as the Vienna Woods, and may be matched by climatic (periglacial) terraces on the nonglacial tributaries (Fink, 1956, p. 71; 1959, p. 50). Here the loess cover on the terraces is thicker, and the soils are less affected by the solifluction of a transitional or subhumid climate. Archeological sites are richest in this region, so that many sections have been studied in detail. The modern soil on loess is Parabraunerde or Braunerde. Precipitation is 25–30 inches (Fink, 1956, p. 52).

East of the Vienna Gates the Danube terrace system breaks down as the river enters the east end of the Alps because tectonic subsidence in the Vienna Basin throughout the Pleistocene caused the older sediments to be buried successively by younger sediments, thus preventing their being preserved as terraces. In this drier region east and north of Vienna and in adjacent parts of Czechoslovakia and Hungary, however, the loess stratigraphy shows soils that are relatively well preserved because of less solifluction, and the loesses are less leached of carbonate and contain more snails and mammal fossils. The modern soil in the dry landscape is Chernozem or (above altitudes of 200 m) Braunerde. Precipitation is less than 25 inches (Fink, 1956, p. 52).

The facies of buried soils in the arid region differ sufficiently from those in the humid and transitional regions that correlation is not always easy. It must be based on general pedologic comparisons, such as color, structure, degree of carbonate leaching and precipitation, preservation of original loess character, etc. Correlation by terrace relations, paleontologic studies, or similar means among the three regions is either not possible or not sufficiently well worked out. The modern and buried soils of the three climatic regions are listed in Table 2.

TABLE 2. CORRELATION OF SOILS
(Adapted from Fink, 1959, p. 55)

	Dry loess landscape north and east of Vienna	Transition region (Krems-Göttweig area)	Humid loess landscape (Linz area and west)
Recent Soil:	Chernozem	Braunerde	Parabraunerde
	Loess	Loess	Loess
	"Stillfried B"	"Paudorf"	"Nassboden" (Tundra soil)
	Loess	Loess	Loess
	Fellabrunn (Stillfried complex):		
	Interbedded loess and humic zones	Humic solifluction loess	Weakly humic solifluction loess
	B horizon of soil	"Göttweig" B horizon of soil	B horizon of Parabraunerde
	Loess	Loess	Loess
		"Krems" B horizon of soil	Pseudogley
		Loess	Loess

KREMS SOIL: The Krems soil at Krems, 30 miles west of Vienna, is found in a thick loess section in which the Göttweig soil occurs at a higher level. The Krems soil here is a complex

of several soils. At the top is a red-brown truncated B horizon of a forest soil about 50 cm thick with a secondary carbonate horizon at the base. This is underlain by several meters of loess with weak red-brown layers that record other phases of soil formation. It was originally correlated by Götzinger (1936, p. 55) as Mindel/Riss but is considered by Brandtner (1956, p. 151) to be Riss/Würm. The complex cannot be certainly identified within the loess section in southern Czechoslovakia where the important mollusc sites are located. The soil called Krems in Czechoslovakia by Brandtner (1956, p. 134–142) is a complex consisting of a reddish-brown clayey B horizon of a forest soil that is truncated at the top and is overlain by two or three chernozemic A horizons intercalated with calcareous loess layers and modified by ice-wedge casts. Whereas the truncated forest soil is considered by Brandtner to represent the R/W interglacial, the overlying parautochthonous chernozemic layers are referred to climatic oscillations at the onset of the Würm cold phase. The complex is developed on periglacial terrace deposits considered to be Saale (Riss) by Prosek and Lozek (1957, p. 75), but the correlation with the type locality of Krems seems by no means certain without additional evidence.

GÖTTWEIG SOIL: The Göttweig soil at the type locality is a truncated B horizon of a forest soil. It is correlated by Brandtner (1956, p. 143; 1954, p. 57) with the Fellabrunn (Stillfried) soil complex found in the drier region of Austria, which consists of a leached and truncated B horizon of a forest soil with a secondary carbonate zone at the base, overlain unconformably by several thin layers of calcareous loess intercalated with partially leached humic soil (chernozemic A horizon) with crotovinas. In the humid and transition climatic regions the humic soil layers of the Fellabrunn complex are represented by humic solifluction loess found at some localities on top of the B horizon.

The description of the Fellabrunn complex resembles closely that stated above for the R/W soil ("Krems") of southern Czechoslovakia (*e.g.*, at Brünn), and in fact it was so correlated by Pelesik, who worked intensively on both types (Brandtner, 1956, p. 146). Brandtner himself, however, believes that the two should not be correlated because they are in different climatic provinces, and their resemblances are accidental—Brünn is in the humid region, and Oberfellabrünn in the arid.

Three radiocarbon dates for humus from the chernozemic layers in the upper part of the soil complex at Oberfellabrunn are in the range 37,000–43,000 years old. These dates fit those set for the Göttweig interstadial by Gross (1958, p. 166), but they apply only to the weak chernozemic humic layers and not necessarily to the main forest soil which unconformably underlies them and which is presumably the equivalent of the Göttweig soil at the type locality. The age of the Göttweig itself is thus not necessarily measured. In fact, Gross's dating of the Göttweig interstadial is based on a series of samples that are not well located with respect to the Göttweig soil. The charcoal sample from Senftenberg, a loess site near Krems, is stated to mark the beginning of the Göttweig interstadial at 48,000 years ago (Gross, 1958, p. 166); actually, however, the sample came from on top of the soil rather than at the base, and furthermore has been redated >54,000 rather than at 48,000 (Fink, 1959, p. 58). Other dates used by Gross for the Göttweig interstadial come from the Upper Paleolithic Willendorf site in loess on the Danube River west of Göttweig, from the Upper Paleolithic Istállóskö cave in Hungary, from England (Upton Warren terrace site), and from Holland. The three samples dated at Willendorf (about 30,000–32,000) come from humic solifluction layers, and the relation to possible buried soils is unclear (Fink, 1959, p. 58). The sample from Istállóskö (30,670) came from a level assigned to the end of the W I/II interstadial on the basis of particle-size analysis of the cave sediments. This cave date and the dates from England (about 42,000) and Holland (28,500–32,000) are not the most appropriate for confirming the stratigraphy of buried soils in Austrian loess.

Just as important as radiocarbon dating in the correlation of the Göttweig soil are the relations to the Alpine glacial-terrace sequence. In the critical region near Krems and Göttweig, along the Danube River valley and its tributaries where the Alpine glaciofluvial terraces are preserved, new field work being completed by Fink and collaborators (kindly demonstrated to the writer in July 1960, to be published in Archaeologia Austriaca 1961) shows stratigraphic and geomorphic relations that require different correlation for the Göttweig soil.

Götzinger (1936, p. 55) believed that the Göttweig soil at the type locality was the Riss-Würm interglacial soil developed partly on "Late Riss" loess, which rests directly on

Riss gravels. Gross reinterpreted the "Late Riss" loess as Würm I and thus correlated the Göttweig as Würm I/II interstadial. The new field studies, however, show that gravel beneath the pre-Göttweig loess in question is not Riss but older, and that laterally in the exposure the Göttweig soil is formed directly on the true Hochterrasse (Riss) and is buried by Würm loess, which contains the Paudorf soil in its upper portion. The Hochterrasse in this region may be distinguished from the Niederterrasse (Würm) below and the Deckenschotter (Mindel and older) terraces above not only by altimetry but also by the degree of erosional modification, and it may be mapped westward up the Danube to the moraines of the Salzach and other Pleistocene glaciers. It is a single unit: it does not have the subdivisinos (Young, Middle, and Old Riss) present in the Württemberg area of southern Germany studied by Weidenbach (1953), where the complications about Jungriss and Mittelriss have arisen.

The relations at Göttweig therefore suggest that the Göttweig soil on the Hochterrasse gravels is R/W interglacial, as Götzinger originally believed. Where it is traced laterally onto a loess parent material in the adjacent upland, it is a truncated forest soil at least as well developed as the modern loess soil of the region and thus deserves an interglacial designation rather than interstadial (Fink, 1959, p. 48). The Niederterrasse bears no buried soils or for that matter no true loess, implying that the Würm meltwater flood plains served as a source for the Würm loess, which covers the Göttweig soil on higher ground, and at the same time the flood plains were active enough to prevent loess accumulation on their own surfaces.

PAUDORF SOIL: If Gross's correlation of the Göttweig soil at the type locality as an early Würm interstadial is thus jeopardized by both radiocarbon dating and field relations, what about the subdivision of the Würm? The Paudorf soil has generally been considered to represent a minor warm oscillation during the epoch of loess deposition. It still shows much of the original loess structure (Brandtner, 1956, p. 145) and is distinguished in the region of the type locality by a flecked reddish-brown and gray and by a crumby structure (Brandtner, 1954, p. 68–73). In the humid region, the Paudorf soil presumably is represented by a "Nassboden," considered to be a type of wet tundra soil characterized by gray mottling and a platy structure (Fink, 1956, p. 58–60; Brunnacker, 1956, p. 44–45). In the dry region the Paudorf soil is well developed, and here it has been dated by C^{14} analyses at Stillfried (about 28,000 years old) and at Unterwisternitz (about 26,000 years just above the Paudorf soil). In the transitional climatic region a new date for the Paudorf near the type locality is 27,570 (Fink, oral communication). Other possible or minimum Paudorf dates from less sure stratigraphic positions (because of solifluction) include those from Pollau (25,000), Aggsbach (25,000), and Willendorf (30,000–32,000).

Conclusions. The case for marked tripartite subdivisions of the Würm in the Austrian loess area is not strong. The so-called Göttweig or early Würm interstadial is founded on the existence of a buried soil. At the type locality this soil may be the Riss/Würm interglacial soil. In the drier region of Austria it is a soil complex of which the upper part may represent weak climatic oscillations about 40,000 years ago and the lower part may be Riss/Würm. The Paudorf soil seems to be correctly interpreted, but it admittedly represents a weak climatic oscillation and may be somewhat older than believed by Gross (28,000 rather than 25,000 years old), thus bringing it so close to the range assigned by Gross to the Göttweig (44,000 to 28,000) that the two would be coincident even if his Göttweig interval were correctly dated.

The loess sections seem to be more nearly complete in Czechoslovakia than in Austria, because the drier climate inhibited contemporaneous solifluction. For the same reason, the mollusk remains are more abundant and provide independent climatic information as well as a correlatable datum in the last interglacial. The Austrian area, however, has the advantage of proximity to the Alpine glaciofluvial terraces to which the loess chronology must eventually be related. It further has had the benefit of radiocarbon dating of some important sites. Eventually radiocarbon dating of mollusc or humus horizons in the Czech loesses may confirm the chronology of Prosek and Lozek (1957), and tracing of the loess-terrace sequence may be possible southward to the Danube glaciofluvial terrace system above the Vienna Gates. Until such time as the Czech chronology is confirmed by radiocarbon dating and the correlation of the soils of the Krems-Göttweig area is more firmly established, the writer believes that the recognition of a major early Würm interstadial in the loess region and in the Alps, and its identification as "Göttweig," should be considered tentative.

These conclusions bring us to a brief consideration of climatic curves for the Late Pleistocene based on other regions. New pollen analyses and radiocarbon dates from deposits of the Weichsel glacial stage in Denmark and Holland give curves (Fig. 9) that show two Early-glacial interstadial fluctuations in which birch-pine forests temporarily returned to that region (Andersen *et al.*, 1960). The base of the sections is firmly founded on Eem (last interglacial) deposits which have characteristic pollen diagrams. The two interstadials are dated at about 64,000 (Amersfoort) and 59,000 (Brørup) years old on the basis of five radiocarbon analyses. No correlative of the Göttweig of Gross is recorded in this region; the time range from 50,000 to 32,000 is assigned in Holland to the Pleni-glacial A, marked by tundra vegetation at best. A possibility of a correlative of the Paudorf interval at about 30,000 years ago is suggested. The following phase, Pleni-glacial B, from about 28,000 years ago until the start of the Late-glacial 14,000 years ago, remains unrecorded by organic sediments in this region.

The curves of late Pleistocene ocean-surface temperatures based on study of planktonic Foraminifera fossils from deep-sea sediments must be mentioned as another measure of European climatic change inasmuch as they presumably reflect the planetary situation. Two types of curve have been constructed, one (Fig. 9) based on oxygen-isotope analysis (Emiliani, 1955), the other on frequency of certain cold- or warm-loving Foraminifera (Ericson and Wollin, 1956). The time scale is based on radiocarbon and ionium dates in the upper part of the cores and extrapolation to depth by estimation of rates of sedimentation (Emiliani, 1955, p. 556 ff.). If the slightly modified time scale of Broecker *et al.* (1958, p. 513) is used, the oxygen-isotope curve shows two cold maxima in the last 70,000 years—the time assigned to the Weichsel by Andersen *et al.* (1960). These two maxima support the bipartite division of the last glaciation used by Gross and others for Europe.

The Milankovitch curve of variations in the intensity of summer solar radiation at 65° N. latitude is also presented as a possible key to late Pleistocene climatic change in Europe (Fig. 9). Such a curve is adopted by Zeuner (1945–1959) as an explanation for the geologic and biogeographic sequence in Europe and is also used by Emiliani (1955, p. 567 ff.) and Fairbridge (1960) for correlation of ocean temperatures and sea-level changes respectively. Emiliani's correlation of the radiation curve with European climatic chronology (Würm I and II cold phases at 72,000 and 23,000 years ago respectively) seems to be more satisfactory than that of Zeuner (WI, II, and III cold phases at 115,000, 72,000, and 25,000 respectively).

PLEISTOCENE CLIMATOLOGY

General

The efforts of Poser (1948; 1950), as reviewed above, to produce maps of Pleistocene winds and mean atmospheric pressure systems for summer and winter represent an approach to Pleistocene climatology that is based on the interpretation of geomorphic features. A more common (but probably more tenuous) approach is based on extrapolation of present-day climatic anomalies under the postulate that mean temperatures were appreciably lower during the Pleistocene.

Attempts at such reconstructions go back at least to Penck (1913) and were quite generalized for many years (Brooks, 1926; Simpson, 1934). They became more detailed after studies on Pleistocene snow line indicated the extent of depression of the temperature, and after studies of the Pleistocene frost features delineated the periglacial zone. Particularly important have been the recent analyses of modern circulation patterns that prevail when Europe experiences anomalous heavy snowfalls, cool summers, or other conditions suggestive of Pleistocene climate.

Pleistocene Temperature Depression

Snow-line depression. The most common method of calculating the Pleistocene depression of temperature comes from the determination of the difference in elevation between the present and the Pleistocene snow lines and the multiplication of this figure by a vertical temperature gradient (normal lapse rate). Pleistocene snow line is generally determined from the lowest elevations of small cirques of firn basins, although an alternative and less accurate method involves the calculation of the elevation midway between the glacier terminus and the crest of the cirque wall.

The history of investigations on Pleistocene snow line has recently been summarized by Morawetz (1955, p. 192–194). Seventy years ago Penck calculated the Pleistocene snow-line depression for the Alps at 1000 m and, assuming a lapse rate of 0.5°C/100 m, arrived at a figure

of 5°C for Pleistocene depression of the mean annual temperature. Klute (1928, p. 80) used this figure in his comprehensive study of snow-line depression in different climatic zones over the whole world. Later, however, it was determined from weather stations in the western British Isles that the difference between the temperature at the modern snow line (estimated to exist just above Ben Nevis in Scotland) and the present temperature at the elevation of the Pleistocene snow line is 8°C, and Penck (1938, p. 87 ff.) accepted this figure as more reliable for Pleistocene temperature depression at least for humid regions.

western United States by Baker (1944, p. 225) gives the following average figures.

January mean	.54°C/100 m
January mean maximum	.54
January mean minimum	.54
July mean	.65
July mean maximum	.74
July mean minimum	.54
Annual mean	*ca.* .6

Large variations from these figures, especially in winter, are attributed by Baker in part to pronounced temperature inversions. The gradient depends on many factors such as frequency

TABLE 3. TEMPERATURE AND SNOW LINE IN THE ALPS
(Weischet, 1954, Fig. 1)

	Mean annual temperature		Pleistocene depression		Vertical temperature gradient
	At elevation of present snow line	At elevation of Pleistocene snow line	Temp.	Snow line	
	°C	°C	°C	m	°C/100 m
Santis	−2.5	+4	6.5	1000	.65
Zugspitze	−2.5	+5	7.5	1100	.68
Sonnblick	−5	+2	7.0	1100	.70

From more recent studies in the Alps we can identify the following facts and opinions. Present snow line ranges from 2500 to 3350 m (Klebelsberg, 1949, p. 660); it is relatively low on the north flank because of the higher latitude, and in the western Alps because of greater snowfall. Pleistocene snow-line depression was 1000–1200 m. Vertical temperature gradients more realistic for ground stations than the traditional free-air rate of 0.5°C/100 m can be determined from a few high-mountain weather stations, so that only slight extrapolations are required to calculate the temperature at the elevation of both the Pleistocene and the modern snow line (Table 3).

Table 3 shows that the mean annual temperature at the modern snow line is not 0°C, as is often assumed, but is generally lower. For the Alps the vertical temperature gradient for ground stations is closer to 0.7°C/100 m than to 0.5°, and in fact this gradient had been used by Köppen (1920) in an early study of Pleistocene depression of tree line. Morawetz (1955, p. 195) cites other modern gradients in the Alps ranging from 0.8°C/100 m at high elevations to 0.3° in the lower valleys. Measurements of the surface lapse rates in 28 mountain areas in the

and character of invading air masses, the local topography, radiational cooling, snow cover, etc., as well as the season. Any major middle-latitude mountain mass such as the Alps can certainly experience a variety of synoptic situations over the years, and the effect of each type on the glacial, geomorphic, and biogeographic relations is perhaps not most reliably expressed by a consideration of annual means or any other means. It is not easy to analyze these climatic elements statistically without the use of means, however, and at present the mean values are the most practical representation of the climatic factors for the discussion at hand.

In a recent thorough study of the climatic conditions in the Sierra Nevada of California, Miller (1955, p. 15–20) discusses the effects of local factors on the vertical temperature gradient, which in this region ranges from .46 to .61°C/100 m for ground stations. Miller also made a comparison of the relation of ground temperatures at high mountain stations in 17 mountain ranges throughout the world to the free-air temperature at the 700-millibar level (elevation *ca.* 10,000 feet) directly above the range. For the comparison, the ground temperatures had been adjusted to a uniform lati-

tude (40°) by the use of sea-level latitudinal temperature gradients, and adjusted to a uniform elevation (7000 feet) by the use of the local surface lapse rate extrapolated to 7000 feet from the closest mountain station available in the particular mountain range. If the normal free-air lapse rate between the 7000-foot ground station and the 10,000-foot air station applies, namely .55°C/100 m (3°F/1000 feet), it would be expected that the temperature at 10,000 feet should be 5°C colder than at 7000 feet. The figures show that the mean temperature for the 3 winter months at 10,000 feet ranges from only 2°C colder to as much as 3°C warmer than at 7,000 feet, and that for spring it ranges from 7°C colder to 1°C warmer. The specific figures for the Alps are a 0°C difference for winter (*i.e.*, isothermal from 7000 feet on the ground to 10,000 feet in the air above) and 3°C colder for spring. The figures illustrate that temperatures at mountain stations probably do not accurately reflect temperatures in the free air beside the mountains at the same elevations, and therefore that the use of free-air lapse rates in predicting ground temperatures in mountain regions is subject to large errors. Use of any lapse rate, whether accurate or not, in the conversion of Pleistocene snow-line depression to temperature depression depends on the further assumption that the Pleistocene lapse rate of the region was the same as the modern. Such an assumption may not be justified in view of the possibly different air-mass and precipitation conditions during the Pleistocene.

Inverted lapse rates. Besides the snow-line studies, other methods of calculating Pleistocene temperature depression have been used—studies of the distribution of fossil plants, animals, and frost features in regions beyond their modern ranges. Iversen (1954, p. 89 ff.), for example, calculated mean July temperatures and other climatic elements on the basis of the modern distribution of plants found as pollen in Late-glacial sediments of Denmark, and Firbas (1949, p. 286–294) made similar estimates for central Europe. Soergel (1936) was the first to calculate temperature depression from frost features (ice wedges) and from the distribution of certain vertebrates. His studies on frost features were followed by those of Poser (1948) as enumerated in the earlier pages of this review.

The basis of Poser's calculation of temperature depression was the intersection of the Pleistocene polar forest border (July 10°C isotherm) and the equatorial limit of permafrost (−2°C annual isotherm) near Zagreb just east of the Alps. The position of the Pleistocene forest boundary was based on the pollen analyses of Firbas and others. Mortensen (1952, p. 153) states that on the basis of more recent studies the permafrost limit marks the −5°C isotherm rather than −2°C, implying a temperature depression in central Germany of 15°C rather than the 12°C calculated by Poser. As stated above, however, sporadic permafrost can occur in regions with mean annual temperatures as high as 0°C (Black, 1954, p. 839). Inasmuch as the Pleistocene frost features recorded by Poser may represent only sporadic permafrost (or even deep seasonal frost), the temperature requirement may thus have been close to 0°C. This possibility, coupled with the potential unreliability of some of the so-called permafrost indicators (as discussed above), makes the precise calculation of temperature depression difficult on the basis of frost features.

It seems probable, however, that the Pleistocene temperature depression close to the ground in the periglacial region of central Europe amounted to at least 10°C, or 2–5°C greater than the figures usually given for the Alps on the basis of snow-line depression. Mortensen (1952, p. 154 ff.) explains the discrepancy by the proposition that the vertical temperature gradient was actually inverted in the Pleistocene at low altitudes, so that the inferred temperature depression was great in the lowlands (where frost features and plant fossils are found) but small in the mountains near the snow line. His evidence for a strong and deep inversion comes from Penck's observation that the Alpine firn basins near the present snow line (2600 m elevation) had no more ice in them during the glacial period than they do at present. He infers from this observation that the temperature and snowfall in the Pleistocene at this elevation must have been the same as today, and that the Pleistocene temperature depression was thus confined to lower elevations. His curve of vertical temperature gradient for the winter shows a sharp inversion below 2500 m. Although his summer curve shows no inversion, a distinct change in gradient at 2500 m is postulated. In this view the Pleistocene depression of the mean annual temperature thus increased from zero at 2500 m to high values (10–15°C) at lower elevations and accounts for the discrepancies between the calculations in the mountains based on snow line and calculations in lowlands based on frost features and on plants and animals.

Mortensen's (1952) inversion hypothesis re-

quires generation of cold air masses over the north European ice sheet and adjacent snow-covered central Europe during the winter, with frequent outbreaks of this cold air to underride warmer air masses and produce the postulated inversion. The atmospheric-pressure pattern worked out by Poser (1950, Fig. 1) on the basis of sand dunes suggests a semipermanent high-pressure system over Europe even in summer—a condition that might be produced by such a cold air mass. Some confirmation of the hypothesis comes from studies of modern lapse rates in arctic continental regions. At Yakutsk in northeastern Siberia the ground inversion in winter depresses the basal air temperature as much as 45°, with the thickness of the inversion layer being about 2500 m (Flohn, 1952b, Fig. 1). The mean January gradients at several North American arctic stations show an inversion up to about 1500 m, which produces surface temperatures 10°–20°C lower than they would be if the upper-air rate continued down to the surface (Meteorological Div., 1944). Although July lapse rates show no inversion, there is a depression of a few degrees. The gradient above the inversion at these North American stations, incidentally, is about 0.6°C/100 m up to about 5000 m altitude, and about 0.7°C from here up to the tropopause at 10,000 m (8000 m in summer).

The winter inversions at high latitudes are a result of radiational cooling under cloudless, windless conditions during the long polar night when the sea is frozen or the ground is snow-covered (Flohn, 1952b, p. 81–82). The winter inversion at Eismitte is 20°C greater in clear weather than in overcast. That daily heating in winter has a negligible effect on the basal temperature is seen from a comparison of noon and midnight readings at Fairbanks, Alaska. The effect of snow cover is shown by the stronger inversion at Eismitte, a station in the middle of the Greenland ice sheet, than at Thule, which is located on exposed land 400 miles to the north.

The cold air generated by ground radiation spreads laterally out from its source region by subsidence (Wexler, 1936, p. 135), and from time to time the entire polar mass breaks outward. Thus a much larger area may be affected by the cold basal air. Pleistocene conditions over the north European ice sheet and a snow-covered central Europe might be less favorable for the development of these conditions in winter than the present arctic regions are, because of the lower latitude involved—the occurrence of daily solar heating in middle latitudes might introduce enough turbulence to retard the development of intense inversions. On the other hand, the pollen-analytical evidence for tundra in central Europe during the glacial period implies that the albedo of the ground surface in winter would be high, because the low vegetation would not project far above the snow cover. The effect discovered by Miller (1955, p. 120–122) in the Sierra Nevada of California would not apply—that even with one of the heaviest snowpacks on the continent the albedo in winter and spring is low because the cover of coniferous trees adsorbs so much of the radiation.

Mortensen's hypothesis, however, assumes that at a station in the Alpine foreland the free-air lapse rate up to a height of 2500 m is the same as the lapse rate on ground stations up the north flank of the adjacent Alps. To the extent that the inversion in the lapse rate is a result of local radiational cooling, however, the height of the inversion would be determined by height above the ground level rather than the height above sea level. Furthermore, the hypothesis assumes that the Alpine mass itself had no effect on the character of the polar air which moved against it. The contrasts between ground temperature and free-air temperature at the same altitude, as cited above from the work of Miller (1955, p. 19), indicate that large mountain masses have important local effects on the temperature of the air nearby.

Flohn (1953, Fig. 2) accepts the probability of a mean inversion caused by ground cooling up to a height of 1000 m. He does not accept Mortensen's hypothesis of no temperature depression above 2500 m, however, but adopts the general figure of 4°C for elevations above the inversion level. This appears to be a more realistic hypothesis, and accounts for the evidence for particularly low temperatures at ground level in the periglacial region of central Europe.

Effects of precipitation. The effects of precipitation on snow-line depression have been examined in different ways. Klute (1928, p. 82 ff.) attributed excessive snow-line depression in certain climatic zones to the lowering of snow line down into the elevation zone of maximum Pleistocene snowfall. The existence of such a zone of maximum snow fall at intermediate elevations at the present time in the Alps, however, has never been proven, according to Klebelsberg (1949, p. 664), and in the northern Sierra Nevada of California, where it has long been assumed on the basis of earlier work, its existence has recently been challenged by

Miller (1955, p. 22) on the basis of new analyses of the snow records.

Firbas (1950, p. 163 ff.), on the other hand, found from pollen studies that the Pleistocene tree line in the Alps was depressed more than the snow line, and attributed this relation to aridity. Vegetation at this time was marked by *Artemisia*, generally considered as an indication of relative aridity. Mortensen (1952, p. 154) points out, however, that these paleobotanical studies were necessarily all from sites at relatively low elevations, where the temperature inversion would have greater effect than at elevations near the snow line. Actually, the loess stratigraphy shows both humid and dry phases, with the climate at the time of maximum glaciation being relatively dry. Weischet (1954, p. 112) suggests the precipitation was 20% less and therefore that the modern temperature gradient in the Alps should not be used without a correction. He therefore believes that a temperature depression of 8–10°C is more accurate for the Alps than 7°C.

In another study of the precipitation effect, Klute (1951) prepared January and July temperature maps for Europe with attention to the local effects of the ice sheet and oceanic influence on the west. The maps are based on Poser's (1948, p. 57) figures for summer and winter mean temperature at the point east of the Alps where the Pleistocene forest border and the permafrost border cross, and contain the revisions of Büdel (1951).

Klein (1953, p. 101–104) has taken the next step. Using Poser's map of climatic zones based largely on frost features (Fig. 3), she estimated the temperature depression for various parts of Europe, ranging from about 9°C in Greece, Italy, and Spain, to 13°C in central Europe and 16°C at the very edge of the ice sheet. Then, using figures for snow-line depression in various mountain regions of Europe, and applying a vertical temperature gradient of 0.9°C/100 m for the continental climatic zone and 0.7°C/100 m for the oceanic zone, she calculated what the temperature depression should be in the several areas under consideration. The discrepancy between this calculated value and the value derived from Poser's frost map is attributed to differences in precipitation with the use of a factor of 150 mm precipitation for each 100 m of extra snow-line depression. For example, the snow-line depression in the Spanish Pyrenees is stated to be 1200 m. The temperature depression for nearby southern France, as inferred from the Poser-Büdel maps, is calculated as 10.6°C, which should be equivalent to a snow-line depression of 1514 m if the humid-region gradient of 0.7°C/100 m is used. The difference 1514–1200 m = 314 m is attributed to an increase in precipitation of (150 mm per 100 m) × 314 m = 470 mm. After computing such discrepancies all over Europe, Klein prepared a map to show the Würm precipitation, and from this Klute (1951, Fig. 3) constructed a map to show the percentage of precipitation during the Würm compared to the present. The percentage map shows high values on the west coasts of the British Isles and in France, Spain, Italy, and the Balkans (60% of today), drier in the Scandinavian mountains and in the Alps, and still drier farther east so that Russia had only 20 per cent of the modern precipitation.

Ingenious though these cartographic representations appear to be, the data on which they are based do not justify their detail. The possible errors in the basic Poser-Büdel maps are great, as are also the difficulties in choosing the correct vertical gradients for conversion of snow-line depression to temperature depression or precipitation change. Errors may thus be compounded. This is not to deny that it was probably drier in eastern than in western Europe during the Würm, but the degree of differentiation cannot be expressed so precisely.

Mediterranean region. For the subtropical mountains a Pleistocene temperature depression of 4°C has usually been assumed on the basis of a snow-line depression of 800 m. Recent studies in these latitudes, however, suggest that the snow-line depression may have rivaled or even exceeded that of the middle latitudes. Thus Mortensen (1957, p. 44–46) refers to studies by different workers that indicate snow-line depression ranging from 1400 to 2100 m in the Sierra Nevada of Spain, the Atlas Mountains, Abyssinia, and the Himalaya, and the present writer has calculated a figure of 1800 m for the Zagros Mountains of Iraq (Wright, 1960, p. 89). These values, when converted to temperature depression on the basis of the generally assumed lapse rate of 0.5°C/100 m, yield figures of 7°–10°C temperature depression for low latitudes—considered to be much too great when applied to low elevations. Mortensen (1957, p. 49) shows that for mountains near Rio de Janiero, where Pleistocene snow line had an elevation of 2300 m, the use of this gradient yields a summer temperature depression of as much as 14°C at sea level.

One way of overcoming this difficulty is by

the postulate that the excess amount of snow-line depression beyond that resulting from temperature may be explained by increased snowfall. Antevs (1954, p. 187) uses such reasoning in a study of Pleistocene snow-line depression in east-central New Mexico (latitude 37°), and a parallel study of a nearby lowland Pleistocene lake indicates an increase in precipitation as well as a decrease in temperature (Leopold, 1951, p. 166).

An alternate explanation for the high values of snow-line depression in low latitudes is the suggestion by Mortensen (1957, p. 50) that the vertical temperature gradient below the Pleistocene snow line is closer to 1.0°C/100 m (the dry-adiabatic rate) than 0.5°C because of the complete instability consequent on the great radiational heating from the ground in subtropical regions. Use of a gradient of about 0.9°C for the Brazilian case would yield a temperature depression of only a few degrees at sea level rather than 14°, permitting the continuance of tropical conditions, whereas at high elevations the temperature was lowered enough to bring about extreme depression of the snow line.

Studies of temperature lapse rates over subtropical continental areas have not been undertaken, but for ocean areas Riehl (1954, p. 52–71) has summarized the observations of the German METEOR expedition in the Atlantic and his own studies in the subtropical eastern Pacific. The most striking feature of these regions is the persistent trade-wind inversion, which is produced by the subsidence of warm dry air from the subtropical high-pressure regions over the cool moist air that originates over the upwelling cold water along the west coasts of Africa and America. In the case of the Atlantic the inversion layer near the source region has a base at an elevation of about 500 m. Its thickness is highly variable but averages about 400 m. The temperature increase upward within the layer amounts to 5°–8°C. The layer persists as the air masses move westward in the tradewind belt, but its intensity weakens, and its base rises to about 1500 m.

As a result of these conditions, lapse rates below the inversion range from superadiabatic (*e.g.*, 1.0° − 1.4°C/100 m) in the lowest few hundred meters to stable (0.5° − 0.7°C) as the inversion base is approached. The counter rate within the inversion is moderate to steep. Above the inversion the air is very dry, and the lapse rate steepens to 0.8° − 0.9°C/100 m.

The extent to which this temperature structure of the air over subtropical oceans applies to the Mediterranean area is problematical. Riehl (1954, p. 50) shows that lapse rates range from stable to inverted for different air masses over Khartoum, Sudan. The continental polar air that occasionally reaches the region in winter has an inversion layer at moderate elevation, attributed to subsidence with divergence.

In view of the extreme variability in subtropical temperature lapse rates thus summarized by Riehl, Mortensen's (1957, p. 50) particular quotations of steep low-altitude lapse rates from isolated localities (*e.g.*, Samoa) are insufficient to support by themselves his explanation for the high values of Pleistocene snow-line depression. His main conclusion, however, seems valid—that snow-line depression alone is insufficient grounds for inferring Pleistocene temperature depression at sea level. His opinion that biologic evidence and frost features are more reliable temperature indicators than snow line cannot be tested for the Mediterranean region until more fossil materials have been found and studied. Although it is hard to believe that all the life zones on subtropical mountains (like the Atlas or Zagros ranges) were depressed an amount comparable to the depression of the snow line, without some telescoping, it will be some time before the pollen-analytical and other paleontological studies of subtropical lowlands and foothills allow the kind of climatic reconstructions that have been made for central Europe.

Despite the conflicting evidence concerning the specific amount of temperature depression in continental regions of the temperate and subtropical latitudes, there seems to be general agreement that the temperature was lowered at least 4°C. Perhaps the most reliable indications for Pleistocene planetary temperatures at sea level, however, may come not from continents but from the study of ocean-bottom sediments. Changes in the frequency of certain temperature-sensitive planktonic Foraminifera in cores of abyssal sediments give a record of temperature fluctuations during sedimentation (Ericson and Wollin, 1956). Changes in the ratio of oxygen isotopes in certain planktonic Foraminifera also reflect surface-water temperatures, and Emiliani (1955, Fig. 15) has constructed a paleotemperature curve that suggests an ocean-water temperature for the glacial phases 6°C lower than the present.

Assuming a general temperature reduction of 4°C and disregarding for the moment the possibility of a general precipitation increase as a

cause for the glacial period, Flohn (1953, p. 270) attempts to calculate the secondary effects of temperature reduction on precipitation. Pointing out that most of the atmospheric moisture comes from the oceans, presumably at low latitudes, Flohn calculates that a 4°C temperature reduction in the Pleistocene should bring about a 20 per cent reduction in evaporation and a comparable decrease in precipitation in this region (0°–5° latitude). For latitudes 20°–25° the reduction of precipitation is postulated as 40 per cent because the evaporation is further decreased by weakening of the trade winds and by southward migration of the trade-wind belt. In central Europe close to the ice sheets the evaporation was especially low because of the even greater depression of the temperature (12°C); relative aridity is suggested for this region during at least part of the glacial phases by the prevalence of loess deposition. Only in the Mediterranean region should the frequency and probably the quantity of precipitation have been greater—the increased frontal activity in this zone could compensate for the effect of lower temperature on evaporation.

General Atmospheric Circulation

Willett (1950) prepared a mean-pressure map for the northern-hemisphere winter during the last cold period; it shows a dominant meridional circulation, with low-pressure cells over the Atlantic and Pacific oceans and glacial anticyclones over southwestern Canada and northeastern Europe. The anticyclones served as blocking highs for the eastward movement of cyclones that originated in the Aleutian and Icelandic lows respectively to the west, and many storms were therefore forced to skirt the highs on the southeast, bringing heavy precipitation to the Gulf of Mexico and the Mediterranean in the winter, and in the summer nourishing the southern margin of the ice sheets. South of the low cells the intensified subtropical high-pressure belt, which today controls the aridity of the Mediterranean, was therefore pushed to the south and restricted, allowing the development of pluvial conditions in North Africa. The tropical easterlies were strengthened as a reflection of the intensified zonal (latitudinal) circulation, and precipitation increased here as well, bringing pluvial conditions to East Africa south of the subtropical dry belt.

Willett based his map in part on the anomalous pressure patterns that prevail today during winter intervals of severe storms, and he would therefore postulate that these anomalous conditions of today were prevalent during the Pleistocene. This type of comparison seems to be a reasonable approach to the reconstruction of Pleistocene atmospheric circulation, and as further analyses of the weather record are made new opportunities become available for the study of conditions favorable to glaciation. Thus Lysgaard's (1949) maps on recent climatic fluctuations and similar studies have been adapted to show that the time of maximum recent glacial advance in 1600–1850 was marked by a pattern of meridional circulation and was preceded and followed by "nonglacial" times with dominant zonal circulation (Willett, 1953, p. 56). Correlation of the "glacial" pattern with times of maximum variations in solar activity is suggestive although not certain, and the resemblance (or lack thereof) between Willett's Pleistocene winter pressure map and Wexler's (1956, Fig. 3) map of January pressure anomalies during sunspot maxima may be noted.

Leighly (1949, p. 141–145) showed that the synoptic situations during the period 1916–1938 that produced heavy snows in interior North America are those with low-index (meridional) circulation. Such a pattern is provided by a high-pressure cell over western or central Canada that not only served as a block to westerly winds but permitted flow of moist air into the Canadian prairie from the southeast. Once the North American ice sheet became extensive it increased the occurrence of these favorable anticyclonic conditions. Leighly admits that the same reasoning cannot be applied to the nourishment of the north European ice sheet, because the moisture source is clearly to the west rather than to the southeast as in North America.

Another example of the application of recorded weather anomalies to Pleistocene climatology is Mather's (1954) study of the mean January pressure, temperature, and precipitation of the 1930's compared to the preceding 30 years. Assuming that the recent warming of the North Atlantic region and the consequent recession of glaciers are related to measurable changes in the general circulation similar in nature to those prevalent in the Pleistocene interglacial phases, Mather's maps of anomalies suggest the following conditions favorable for Pleistocene glaciation:

(1) The Icelandic and Aleutian lows were

displaced to the west compared to today, and the Siberian high to the east. For Europe this meant reduction of the latitudinal pressure gradient and consequently a decrease in the intensity of southerly air flow over northwest Europe.

(2) Areas of past continental glaciation were colder and drier than today, and unglaciated continental areas were warmer and wetter. The different patterns require that the cause for the climatic changes have effects on continents different from effects on oceans. Changes in solar radiation, either in intensity or in wave length, produce differential effects of this type.

(3) Storm track were shifted southward in North America as a result of the westward shift of the Aleutian low. Temperature was lower in western Canada, and precipitation was higher in the Great Basin of the United States.

Still another study of recent weather changes (Lamb and Johnson, 1959) is the analysis of January atmospheric pressure (the key to the general circulation) the world over for each decade from 1760, and the statistical correlation of pressure changes with temperature, precipitation, ocean currents, and the extent of the arctic ice pack. The authors do not speculate on the extrapolation of these recent changes to the Pleistocene but suggest that Willett's concept of expansion and contraction of the circumpolar vortex is oversimplified because it ignores the asymmetry of development of the two principal ice sheets and other geographic factors.

Flohn (1952a; 1953) essentially follows Willett's picture of Pleistocene circulation and adds more details based in part on analogies with selected anomalous patterns of modern severe winters. The mean upper-air maps for the present show cold poles over northern Eurasia and northern Canada. The Canadian center provides a frontal zone against moist air masses of the western Atlantic, resulting in nourishment for the Greenland ice sheet. During severe winters these cold centers and their associated upper-air troughs shift to the west and become stronger. If in the Pleistocene this were the prevailing condition, along with increased meridional circulation, then the polar front would be expanded in activity, and the Labradorean and Scandinavian ice sheets would grow to the southwest as well as to the southeast of the respective cold centers. That the southward and westward expansion of these high-latitude blocking highs can have effects even on the lower latitudes is seen from the occurrence of heavy precipitation in the dry Great Basin of the western United States in January 1949, related to this pattern (Flohn, 1952a, p. 169).

For the Mediterranean, Flohn (1952a, p. 162) shows that in modern severe winters the cold air penetrates gaps in the weakened subtropical high, bringing rains to the dry belt and increased cyclogenesis in the tropics. A similar condition may have dominated the Pleistocene, although Flohn differs from Willett in presuming that the subtropical high was weakened rather than intensified, allowing pluvial conditions here that are certainly correlated with the glacial phases of Europe (Büdel, 1952).

In a further attempt to relate the present to the past for the Mediterranean climate, Butzer (1957a) has found that the mean atmospheric pressure was appreciably higher during 1921–1940 for both winter and summer in the Mediterranean than in preceding decades, and that the precipitation (years 1911–1940) was lower. This relation is explained in terms of Flohn's (1953, p. 267) concept of intensification and northward shift of the subtropical high during an interpluvial (interglacial) phase, northward shift of the ploar front and storm tracks in the belt of westerlies, and decreased cyclogenesis in the tropical convergence zone resulting in fewer outbreaks of polar air through the strengthened subtropical high.

In another analysis Butzer (1957b) points out that according to the geologic evidence the temperature fluctuations of the Pleistocene in Europe were not necessarily in phase with the changes in precipitation, and therefore that the changes in the general circulation of the atmosphere may have been more complicated than generally assumed in the reconstructions of Willett and Flohn. The loess sequence, for example, indicates that each cold phase started with a relatively moist interval, marked by the formation of solifluction loess, and then continued with a relatively dry interval characterized by deposition of pure loess without solifluction (Büdel, 1951, p. 25). Whether this sequence marks each of the major subdivisions of the Pleistocene (*e.g.*, Würm), as Büdel assumed, or the minor subdivisions as well (*e.g.*, Early Würm, Main Würm), as is implied by the climatic curves of Gross, Woldstedt, and others, depends on one's concept of the scope and correlation of the Würm. Butzer has attempted to cite geologic evidence from the Mediterranean region pointing to a similar kind

of precipitation fluctuation for each phase of the Würm as well as a general decrease in humidity throughout the Würm, but the paleoclimatic interpretation of cave deposits and other geologic and paleontologic features of the Mediterranean region, as well as their detailed correlation with possible subdivisions of the Würm sequence of Europe, is still uncertain (Wright, 1960, p. 87).

With this assumption that each cold phase (both major and minor) was marked by a cool moist portion (pluvial in the Mediterranean) followed by a cold dry portion (with maximum of glaciation), Butzer (1957b) proposes that the principal change in the general circulation to bring about a glacial-pluvial epoch involved primarily precipitation rather than temperature. He therefore supports the Simpson-Willett hypothesis that Pleistocene circulation changes were not the secondary effect either of a change in planetary temperature or of the spread of ice sheets, but instead were a result of changes in solar radiation as they affected the intensity of the circulation. The moist cool phase that was brought about by an increase in solar radiation was marked by increased meridional circulation, southward shift of the polar front, and weakening and penetration of the subtropical high and trade-wind belt as explained by Flohn (1952a, p. 161). The second phase, with a cold-continental aspect and maximum glaciation in Europe, is not explained in detail by Butzer, who appeals to complications brought about by the presence of the ice sheets. On this basis one could make a case that the maximum coldness of the second part of each phase was in fact a secondary result of growth of the ice sheets, which had been started in their advance by the humidity of the first part of the phase and reached their maximum only after long retardation. Butzer concluded that the reversal from dominant meridional circulation to latitudinal circulation was synchronous with the change from moist to dry climate within the cold period, rather than with the temperature change that brought about the end of the cold period.

Change in Solar Radiation as a Cause for Climatic Change

Possible causes for Pleistocene climatic fluctuations may be grouped as terrestrial causes related to changes in topography, volcanic activity, etc., or as solar causes related to changes in the amount of solar radiation received by the earth at various latitudes. The effects of topography on ocean currents and other elements of climatic control (Brooks, 1926, p. 302–320) cannot be ignored and in fact have recently been re-emphasized in a new theory of climatic change by Ewing and Donn (1956). The difficulties in accounting for the cyclic nature of Pleistocene and Recent climatic changes by cyclic changes in topography, however, have made the solar hypotheses more popular in recent years (*e.g.*, Flint, 1957, p. 495–503). Cyclic solar changes of short duration are known from the occurrence of sunspots, and much interest has been directed toward determining the relations of the weather and climate to sunspot cycles and possible longer cycles of similar character.

It might be assumed that a decrease in incoming solar radiation should result in cooling of the atmosphere and bringing about glaciation. Simpson (1934, p. 428 ff.), however, pointed out that decrease in radiation would have the greatest effect in tropical latitudes and therefore would result in a decrease in the poleward temperature gradient. As a result, the intensity of circulation and the consequent storminess that he believed necessary for glacier growth would be lessened. Increased radiation, on the other hand, would increase the circulation and storminess; the intensified cloud cover in the middle latitude storm belts would insulate the ground from the increased solar heating, and the additional snowfall would build the ice sheets at favorable latitudes. The ice sheets would grow, favored by the self-intensifying effects of the glacial anticyclone, until such time as the equatorward extent of the ice and the still increasing solar radiation caused excess melting of the ice and thus retreat. Then would follow a warm wet interglacial phase during the maximum of the radiation curve. The next glaciation should occur during the descent of the radiation curve, leading to a cool dry interglacial.

Simpson's (1957, p. 469 ff.) latest defense of his hypothesis presents a correlation of the interglacial phases of middle latitudes with the pluvial phases of South Africa rather than with interpluvials, as is usually assumed. His calculations suggest temperatures in the periglacial region only 1°–3°C lower than the present rather than the 8°–12°C usually inferred from geomorphic and paleontologic evidence. For latitudes below 45°, the temperatures calculated are as much as 6°C higher than today. Simpson must therefore depend on the secondary refrigeration effects of the ice sheets to

produce the actual cooling recorded in the periglacial region. Geologic support for the hypothesis is generally weak, and its principal merits seem to be in climatologic theory.

Willett (1953, p. 61 ff.) accepts the Simpson hypothesis with enthusiasm but has modified it by suggesting that the radiation changes were confined to the ultra-violet range or to the emission of charged particles (as in the sunspot cycle), permitting changes in circulation intensity without necessarily much heating. Willett (1949) believes that the short-period changes in circulation patterns as analyzed from weather records show relations to changes in solar activity as recorded by sunspot cycles; he suggests that long-period climatic fluctuations (including those of the Pleistocene) resemble the short-period changes so much that they also must be controlled by radiation cycles, of larger magnitude and duration. He shows a correlation between storminess, glacier recession, and groups of four consecutive active sunspot maxima—the time from 1917 to the present is an example (Willett, 1951). This correlation, however, seems to controvert rather than support the hypothesis, which requires decreased rather than increased solar radiation for glacier recession. Wexler (1953, p. 84), on the other hand, shows a correlation between individual sunspot maxima and the winter development of high-pressure cells and of excess precipitation at high latitudes—conditions conducive to glacier expansion. He then makes the following disarming statement, however (p. 84):

"There is as yet no proof that these particular pressure, temperature, and precipitation difference patterns are uniquely determined by taking differences between sunspot maxima and minima. Nor can such proof be readily produced. It may well be that similar patterns can be obtained by selecting comparable groups of years at random, without any regard to sunspot number, and taking differences. When, in fact, such a random experiment was performed for the winter and summer half-years, the patterns and magnitudes of pressure differences were remarkably similar to those of sunspot maxima minus sunspot minima found earlier."

In a later paper, however, Wexler (1956, p. 481–488) abandons the caution and makes a stronger case for the relation between sunspot number and glacier expansion. His maps of winter pressure, temperature, and precipitation anomalies during the maxima of the last four sunspot cycles show conditions favorable for glacier growth in eastern Canada and northern Europe. The case is different for other regions, however, and Wexler's maps show conditions unfavorable for glaciation in southeastern Alaska during these same times of sunspot maxima. This serves as an explanation for Lawrence's (1950, p. 217) discovery that recent minor moraines formed by slight glacier advances during the past 200 years may be correlated with sunspot minima rather than maxima, as worked out on the basis of tree-ring studies. Extension of this explanation to the Pleistocene may not be justified, however, for all the evidence from elsewhere in Alaska suggests general synchroneity of glacier advances here with those of the ice sheets in the Great Lakes region and northern Europe (Karlstrom, 1957).

Although the relations between sunspots and large-scale pressure and weather patterns may hardly be considered established, "there are enough promising clues, both theoretical and observational, to keep at a high pitch the interest of the investigator" (Wexler, 1953, p. 84). One of the weak points of the correlation is that the summer anomalies for sunspot maxima show conditions favorable for ablation (warm, dry, cloudless) rather than for preservation of the winter snow cover. Wexler (1956, p. 488–494) therefore essentially discards the high-radiation hypothesis and instead favors a low-radiation hypothesis based on the periodic blanketing of the atmosphere by volcanic dust. Postulating a 20 per cent reduction in radiation, which would have a greater effect over continents than over oceans, Wexler calculates that a pressure pattern in winter would develop that would result in heavy snowfall in northeastern North America and in Norway. Equally important, the summer temperatures would be cool enough to inhibit melting of winter snow. Wexler points out that both this hypothesis and the sunspot hypothesis depend on ocean-continent contrasts, so the situation should be different in the southern hemisphere.

The natural skepticism about the paradox "hotter sun, cooler earth", inherent in the Simpson-Willett correlation of glaciation with increased solar radiation, has led to the representation of the opposite view by Viete (1950), who makes a case for the correlation of glaciation with *decreased* solar radiation. This hypothesis assumes that the initial reduction in temperature at high latitudes would result in expansion of the polar sea ice and sufficient cooling of adjacent land areas so that ice sheets would grow. Despite the initial decrease in poleward temperature gradient, circulation in-

tensity, and storminess—resulting from the fact that radiation changes have greater effect at low latitudes than at high—Viete suggests that the increased latitudinal circulation resulting from greater contrasts in land and sea temperatures would produce the local storminess necessary for glacier nourishment. Once the glaciers reached a moderate size they would have secondary cooling effects for further expansion, in the manner emphasized by Brooks (1926, Ch. 1). As the ice sheets reached lower latitudes the meridional temperature gradient would eventually be increased to the point at which the increased circulation and storminess would bring about the precipitation required for further ice-sheet nourishment.

The case for strong secondary cooling and self-accelerating glacier growth, however, is not so strong as when it was emphasized by Brooks (1926, Ch. 1) as the major control on glacier extension. Pertinent here are the arguments of Büdel (1953, p. 251) in his opposition to the use of the term "periglacial", that the European Würm climate was not a reflection of glacier expansion but was instead a direct result of the basic climatic change and was only locally modified by the presence of the ice sheet. The most recent climatic curves for the Würm of Europe, as based on the geologic and paleontologic evidence (Woldstedt, 1958, p. 244; Andersen *et al.*, 1960), indicate a long era of low temperature before the glaciers reached their maximum extent near the end of the Würm, but cause and effect cannot easily be separated here. Willett (1950, p. 179) in fact admits that "the question is left unanswered as to the degree to which the major climatic changes of an ice age are a preceding or a contemporary cause of glaciation on the one hand, or a contemporary or subsequent result, on the other hand". He is therefore forced to base his opposition to Viete's argument on climatological theory by insisting that ice-sheet growth can come about only through the vigor of an increased meridional circulation and of an accelerated condensation cycle as impelled by increased solar radiation.

Summary and Conclusions

Lowering of temperature during the last glacial phase of the Pleistocene has been calculated from the depression of the snow line in mountains or from the distribution of cold-loving plants and animals and of frost-disturbed soils. The results, as reviewed in part in recent years by Morawetz (1955), Mortensen (1952; 1957), Klein (1953), and Weischet (1954), among others, are inconsistent and range from 3° to 15° for different climatic regions according to different authors. The wide range of variation may be attributed primarily to insufficient knowledge of the vertical temperature gradient (lapse rate) for the Pleistocene by which the snow-line depression is converted to temperature depression.

Frost soils and vegetation, two criteria for regional temperature, may actually be controlled in part by limiting conditions of severe winter temperatures or length of the frost season, and these in turn may be controlled by ground cooling of sufficient magnitude and frequency that a mean inversion in the lapse rate, at least in winter, might result. Frost soils may well be confined to certain favorable localities and geologic conditions, and their distribution may be the result simply of more frequent occurrence of ground-radiational cooling. They may thus give a distorted picture of the temperature depression that actually prevailed at a greater distance above the ground. The distribution of tundra plants is perhaps a more reliable criterion of regional climate if it is based on numerous pollen diagrams that reflect upland as well as local lowland vegetation, but again it gives information only about the climate near the ground. Lapse rates in the lower few thousand meters of the present atmosphere range from 1.0°C/100 m down to even a negative gradient (inversion) during winter. The relations between the observed gradient and the climatic region, season, elevation zone, and other elements are poorly known, however, and a careful study of these factors should be made before the appropriate gradient might be postulated for Pleistocene conditions. The effects of possible Pleistocene changes in precipitation on snow line are also difficult to evaluate.

Use of the distribution of biologic and morphologic features as Pleistocene temperature indicators is also subject to error. The best Pleistocene vegetational reconstructions are based on pollen analyses (Firbas, 1939; 1950), but frequently the pollen diagrams cannot be dated well enough to establish correlation with a particular climatic phase under consideration. Also, precipitation changes may have an added effect on plant distribution. Animals reflect the vegetational types and are thus one step further removed from the climatic factors. Frost features, as mapped especially by Poser (1948), point to low temperature, but only a few are

reliable indicators of permafrost, and observations of present permafrost show that the temperature required for permafrost is controlled as well by many factors of topography, geology, and hydrology.

This situation leaves the skeptic with grounds for ignoring the entire procedure of searching for a geologic measure of the extent of temperature depression in the Pleistocene. Yet the climatologist must depend on the best geologic judgment as a base for attempts to reconstruct the atmospheric circulation patterns for the Pleistocene. For example, one of the leading solar-radiation theories for Pleistocene climatic changes—that recently defended by Viete (1950)—is based on an assumption of significantly lower planetary temperatures during the glacial phases, whereas the other (Simpson-Willett) depends on increase in precipitation despite slightly *higher* temperatures. The direction and magnitude of the Pleistocene temperature changes should therefore be determined from geologic evidence because they are vital to these and other theories.

The geologic and biogeographic evidence reviewed indicates that temperatures close to the ground in Europe were much lower during the glacial phases than they are today, and that the climate of central Europe was probably drier, at least during parts of the glacial phases. The ocean-sediment studies suggest that planetary temperatures were also lower. These conclusions do not support the Simpson-Willett hypothesis that calls for increase in temperature and precipitation to bring about glaciation in the middle latitudes.

Where not controlled by such geologic or biologic evidence for Pleistocene temperature or precipitation values, climatologic theory as applied to any detailed reconstructions becomes highly speculative. Potentially the most productive approach is probably the study of those recent anomalies in pressure, temperature, and precipitation that appear to be favorable for heavy winter snowfall and low summer temperatures in the once-glaciated regions. Four recent statistically based analyses of such anomalies are cited in the present review as examples—Leighly's (1949) study of recent snowy winters in interior North America, Mather's (1954) study of the patterns of the 1930's, Butzer's (1957a) analysis of recent trends in the Mediterranean, and Wexler's (1956) investigation of the situation during recent sunspot maxima. Beyond these, the comprehensive discussions of Pleistocene climate by Willett (1949; 1950; 1953), Flohn (1952a; 1953), Viete (1950), Simpson (1934; 1957), Brooks (1926–1949), and others are largely theoretical and speculative, and it will be some time before the cause and course of Pleistocene climatic change are established in their main details.

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Short Notes

FRED A. DONATH

EXPERIMENTAL STUDY OF SHEAR FAILURE IN ANISOTROPIC ROCKS

Abstract: Preliminary work has shown that planar anisotropy (foliation) may have a marked effect on both the breaking strength and the angle of shear fracture in rocks. For rocks experimentally deformed at room temperature and under low confining pressure, curves of breaking strength versus inclination of anisotropy are concave upward and parabolic in form. Shear fractures tend to develop parallel to well-developed planar anisotropy for inclinations up to 45°–60° to the direction of maximum pressure.

Introduction

An experimental study is being made to determine the effect of primary and secondary foliation on the strength and failure characteristics of certain common rocks. Several rocks with different types of planar anisotropy (bedding, cleavage, and schistosity) are being tested at room temperature under successively higher confining pressures to determine the effects of the anisotropy during moderately brittle, moderately ductile, and ductile behavior.

The results of preliminary studies at low confining pressures indicate that planar anisotropy has a marked effect both on breaking strength and on the angle of shear fracture. The lowest confining pressure used in the preliminary studies was 30 bars; the highest was 500 bars. Under these pressures, representing depth equivalents in the earth of about 360 to 6000 feet, all rocks tested showed moderately brittle behavior, and failure was by fracture.

Bulk material was collected parallel to foliation wherever possible so that the same layers would be present in all specimens of a given rock type. The bulk material was sawed into rectangular blocks, and 1-inch diameter cores were cut at angles of 0°, 15°, 30°, 45°, 60°, 75°, and 90° to the plane of anisotropy. Cores were trimmed to 2½-inch lengths, placed between a piston and anvil, jacketed, and inserted in a pressure vessel. Axial load was produced by placing the pressure vessel between the platens of a hydraulic press; confining pressure was produced by means of a Blackhawk pump, with kerosene as the confining medium. Each orientation was tested at least twice at a given confining pressure to check reproducibility of data.

Results obtained for the Martinsburg slate (Ordovician; Bangor, Pennsylvania) are presented here, as they illustrate the effect of well-developed anisotropy (cleavage) during moderately brittle behavior of the rock. Data presented are from compression tests. The differential stress at failure represents the breaking strength of the rock under the imposed conditions of the test and is equal to the axial stress minus the confining pressure.

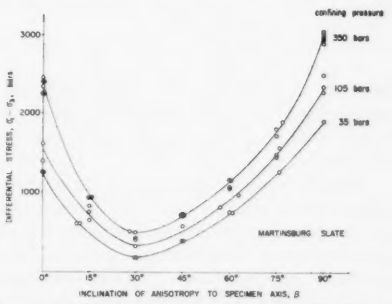

Figure 1. Breaking strength *vs.* inclination of cleavage, Martinsburg slate

Experimental Results

Curves of differential stress at rupture versus inclination of anisotropy are concave upward and parabolic in form. Figure 1 shows the

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results from several runs on Martinsburg slate at three different confining pressures. In slate, cores cut at 90° to the cleavage show the greatest breaking strength—approximately 3000 bars at 350 bars confining pressure; cores cut at 30° show the least breaking strength. Increased confining pressure produces a noticeable upward shift of the curves, as would be expected.

Curves for other rock types tested also tended to be concave upward and roughly parabolic in form; however, the shape and definition of the curves were affected appreciably by the nature of the anisotropy and by the ductility of the rocks under the conditions imposed.

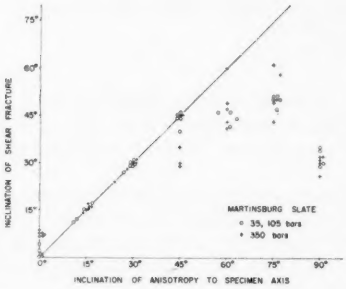

Figure 2. Effect of cleavage on the angle of shear fracture, Martinsburg slate

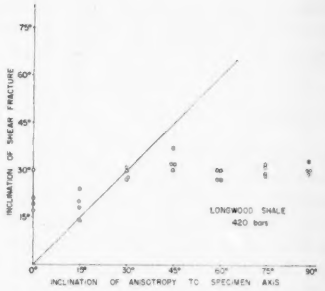

Figure 3. Effect of bedding fissility on the angle of shear fracture, Longwood shale

Experimental data and field observations indicate that in macroscopically homogeneous and isotropic rocks shear fractures develop parallel to the intermediate principal stress and at an angle of about 30° to the maximum principal stress, as predicted by theory. In anisotropic rocks the angle of shear fracture varies considerably; it is dependent on the inclination of the anisotropy with respect to the principal stress directions.

Figure 2 gives the results from several runs on Martinsburg slate for three different confining pressures. Data points along the line indicate that shear fracture developed parallel to cleavage. This is seen to be the case for the 15-, 30-, and some 45-degree orientations. Cleavage oriented parallel to the specimen axis lowers appreciably the 30-degree angle that would be predicted for isotropic materials. Some effect of cleavage appears to be present even in the 60- and 75-degree orientations, but for cleavage oriented at 90° the fractures are inclined at about 30° to the axis of loading. In all except the 90-degree orientations the "strike" of the shear fracture was parallel to that of the cleavage.

Increased confining pressure tends to reduce the effect of anisotropy on shear fracture, as suggested by the 45-degree orientations of specimens of Martinsburg slate. At confining pressures of 35 and 105 bars, shear fracture developed parallel to cleavage. However, at 350 bars confining pressure, shears generally developed at smaller inclinations to the axis of loading. Early results on several other rock types (Donath and Cohen, 1960) indicated that anisotropy had no effect on shear fracture for inclinations greater than 30°. Figure 3 is representative of the results obtained for tests on shale and limestone. For Longwood shale (Silurian; Highland Mills, New York) subjected to 420 bars confining pressure, shear fractures developed at about 30° for nearly all orientations of bedding. The bedding appears to affect the angle of shear fracture only at orientations of 30° or less, and then not too strongly. These results do not reflect a poor development of anisotropy in the rock. The Longwood shale has a well-developed bedding-plane fissility produced by the parallel arrangement of mica flakes and elongated quartz particles. The rock tends to break easily along bedding planes. The lack of influence of anisotropy on shear fracture in this case appears to reflect the ductility of the material being deformed; the ductility is a function of both rock type and confining pressure.

Poorly defined anisotropy, such as indistinct and irregular bedding, did not have the effect on shear fracture that cleavage or schistosity

did. Shear fractures commonly followed both cleavage and schistosity for inclinations up to 45° or slightly greater, particularly at low confining pressures, but many shear fractures seemed to be independent of poorly defined bedding. In the Manlius limestone (Silurian; Alsen, New York) for example, quite irregular variation occurred—from 15° to 45°—at 150 bars confining pressure; at 350 bars pressure, however, shear fracture developed at 25°–30° for nearly all inclinations of bedding.

Data were compared for all tests on Martinsburg slate in which shear fracture developed parallel to cleavage in order to determine the criterion for failure in the plane of cleavage (Fig. 4). Each point on Figure 4 represents the value of normal stress and shear stress that existed at failure on each shear fracture parallel to cleavage. The points fall along a straight line, which presumably represents the ratio of shear stress to normal stress that, if it exists on any cleavage plane, will result in failure along the cleavage. The equation for the line, as determined by the method of least squares, is

$$\tau = 89.8 + 0.329\,\sigma\,. \qquad (1)$$

The angle of internal friction in the plane of cleavage is thus determined to be 18°, and the average cohesive resistance along the cleavage is 89.8 bars. The scatter or standard error of estimate is 36.7 bars, and the cohesive resistance may therefore be expected to range from about 50 to 125 bars along different cleavage planes.

The empirical criterion for failure along the cleavage seems well substantiated by the experimental data and supports the Coulomb theory as discussed by Jaeger (1960).

Mohr circles for a representative run on Martinsburg slate at 350 bars confining pressure are shown in Figure 5 along with the line (1) representing the criterion for failure parallel to cleavage. The circles give the states of stress at failure for the respective orientations of cleavage, and the points on these circles indicate the normal stress and shear stress existing on the cleavage for each orientation. For inclinations of 15°, 30°, 45°, and 60°, the points lie on or very close to the line representing failure along the cleavage. Failure actually did occur parallel to cleavage for these particular tests at 15°, 30°, and 45°. Shear fracture occurred at 49° in the 60-degree specimen; apparently the cohesive resistance along the cleavage for this particular specimen was somewhat larger than the average value indicated in Figure 5. No shear stress exists on planes parallel and perpendicular to the maximum principal stress, and shear failure along the cleavage for these orientations—*i.e.*, 0° and 90°—would not be expected. The ratio of shear stress to normal stress along cleavage oriented at 75° is far too low to favor failure parallel to cleavage.

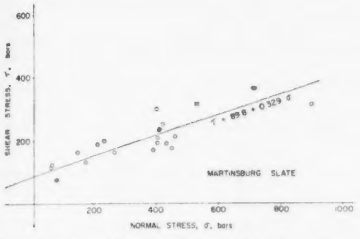

Figure 4. Criterion for failure in the plane of cleavage, Martinsburg slate

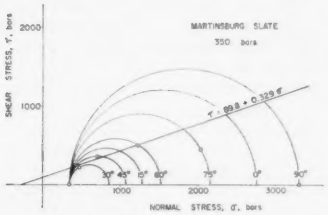

Figure 5. Mohr circles at failure for a typical run of Martinsburg slate with the criterion for failure in the plane of cleavage

A simple representation of two shear strengths, one along the cleavage and another for failure inclined to cleavage, is not adequate to explain failure in Martinsburg slate. In addition to the relatively low strength along the cleavage, represented by equation (1), the strength of Martinsburg slate is different for different directions through the rock. This is apparent from Figure 5. Failure occurred in specimens of 0°, 75°, and 90° orientation, subjected to the same confining pressure, under quite different maximum principal stresses.

Jaeger (1960) has treated the possible effects of anisotropy for a hypothetical case in which the cohesive resistance is continuously variable according to the relation

$$\tau_0 = a - b \cos 2(\alpha - \beta), \qquad (2)$$

where τ_0 is the cohesive resistance in the direction α to the maximum compression, a and

b are constants, and β is the inclination of anisotropy (direction of least cohesive resistance).

If a Coulomb criterion is assumed, the criterion for failure becomes

$$(\tau_{mx} + b \sin 2\beta) \sin 2\alpha + (\mu\tau_{mx} + b \cos 2\beta) \cos 2\alpha = a + \mu\sigma_m \quad (3)$$

for the above condition, where τ_{mx} is the maximum shear stress, σ_m is the mean pressure, and μ is the coefficient of internal friction. This criterion may be written in terms of the maximum and minimum principal stresses, σ_1 and σ_3, respectively, as

$$\sigma_1 = \frac{\sigma_3 \cos \alpha (\sin \alpha + \mu \cos \alpha) - b \cos 2(\alpha - \beta) + a}{\sin \alpha (\cos \alpha - \mu \sin \alpha)} \quad . \quad (4)$$

Data thus far obtained indicate that the coefficient of internal friction in Martinsburg slate is generally close to 0.5 for most orientations of cleavage and that with orientation the cohesive resistance may range from about 5 bars to about 720 bars. These values were determined by plotting Mohr circles for failure and drawing curves tangent to the circles for each of the respective orientations. The slopes of the tangential curves represent the coefficients of internal friction, and the intercepts on the ordinate axis represent the cohesive resistance (Hubbert, 1951). The results for Martinsburg slate are summarized in Table 1.

For $\mu = 0.5, \alpha = 30°$, and confining pressure $\sigma_3 = 350$ bars, the equation of the general type (4) which best fits observed data is

$$\sigma_1 = 236 - 149 \cos 2(30° - \beta). \quad (5)$$

The curve (5) is shown in Figure 6 with the experimental results for 350 bars confining pressure. Very good agreement exists for all orientations except $\beta = 0°$; agreement is probably lacking for this orientation because failure by simple shear fracture was not possible, owing to the restraining effect of the piston and anvil.

The values of the constants a and b, as determined from equations (4) and (5), are 445 and 459, respectively, and equation (2) thus becomes

$$\tau_0 = 445 - 459 \cos 2(30° - \beta). \quad (6)$$

The variation with orientation of cohesive resistance (6) is shown in Figure 7 with the "best" values (points) and possible range (lines) of τ_0 indicated by present data. Good agreement is observed for most orientations.

Table 1. Coefficient of Internal Friction and Cohesive Resistance as a Function of Cleavage Orientation in Martinsburg Slate

Asterisks indicate values by the best Mohr envelopes for existing data; the possible range of values are given in the other two columns.

Orientation	μ	μ^*	τ_0, bars	τ_0^*, bars
0°	0.742 - 0.820	0.820	220–340	300
15°	0.370 - 0.593	0.445	150–240	150
30°	0.310 - 0.534	0.534	5–100	40
45°	0.350 - 0.512	0.512	50–140	120
60°	0.425 - 0.660	0.530	200–280	240
75°	0.430 - 0.595	0.538	340–480	420
90°	0.485 - 0.810	0.622	540–720	630

Geologic Implications

The geometry of rock deformation is the basis for analysis of all geologic structure; therefore, any factor that seriously affects the geometry of deformation may similarly affect the validity of the analysis. Many methods of structural analysis, theoretical and empirical, treat the deformed rock as an initially homogeneous and isotropic mass. Preliminary work has shown that anisotropy may have a striking effect on the deformational geometry of a rock showing moderately brittle behavior.

In faulted rocks, even if the direction of initial slip and the relative sense of slip along a fault were known, the probable direction of maximum principal stress responsible for the fault could not be determined with certainty from this information alone, if the fault were parallel to foliation. In a rock having a distinct planar anisotropism and deformed while moderately brittle, a shear fracture might develop at any angle up to perhaps 60° to the direction of maximum principal stress. This

possibility should not be construed as an argument against Coulomb's law (often referred to as Hartmann's law); planar anisotropy may account for a large inclination of a single shear fracture to the direction of maximum compression, but it does not favor the development of a conjugate shear at a similarly large inclination.

For rocks of similar physical character undergoing folding, slip between layers might be expected to be a controlling mechanism during moderately brittle behavior until fold limbs are inclined at 45°–60° to the direction of maximum principal stress. For greater inclinations, another mechanism must take over—either flow within layers or flow or slip across layer boundaries. The mechanism that would operate would be dependent on the behavior of the rocks which, in turn, is dependent on the rock type and the environmental conditions.

Although the results reported here deal with moderately brittle behavior, similar effects will probably be observed for moderately ductile and ductile behavior. For example, flow might be confined to specific layers under certain conditions, but may transgress layer boundaries under other conditions. The geometry of flow could be appreciably affected by the presence of anisotropy.

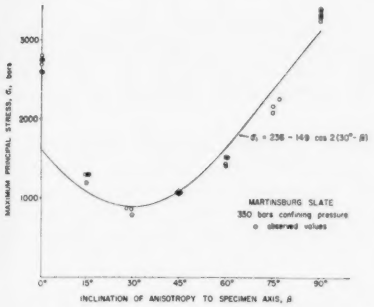

Figure 6. Maximum stress at failure as a function of β, Martinsburg slate

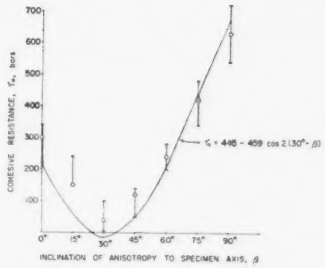

Figure 7. Cohesive resistance as a function of β, Martinsburg slate

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CONTINENTAL GLACIATION IN RELATION TO McFARLAN'S SEA-LEVEL CURVES FOR LOUISIANA

Abstract: McFarlan's sea-level curve for Louisiana is compared with an interpretative glacial-advance curve for the Wisconsinan of the Lake Michigan glacial lobe. A significant ice withdrawal during Farmdalian time is not recorded in the sea-level curve. World-wide glacial fluctuations should be reflected in sea-level variations when sufficient data become available from both environments.

McFarlan (1961) has recently presented curves depicting the position of sea level in southern Louisiana during Wisconsinan time. His calculated points were placed in time by a well-documented group of radio-carbon dates. Because the late Pleistocene fluctuations of sea level are judged to be controlled largely by the accumulation and dissipation of glaciers on the continental areas, the relationship ought to appear in similarity between a curve showing sea-level fluctuation and one depicting the world-wide advance and retreat of continental glaciers. In an attempt to examine this relationship for one area of continental glaciation, we have plotted, in Figure 1, such a curve for the Wisconsinan glaciers of central North America superposed on a reproduction of McFarlan's sea-level curves.

The curve of glacial advance and retreat is based largely on the one presented by Frye and Willman (1960) for the Lake Michigan glacial lobe. In Illinois and Wisconsin finite radiocarbon dates cover the time span back to 37,000 radiocarbon years B.P. These dates have been supplemented by recent publication of dates ranging back to 47,500 radiocarbon years B.P. from early Wisconsinan interstadial deposits in southern Ontario, Canada (deVries and Dreimanis, 1960). Although the Ontario dates are not from the Lake Michigan lobe,

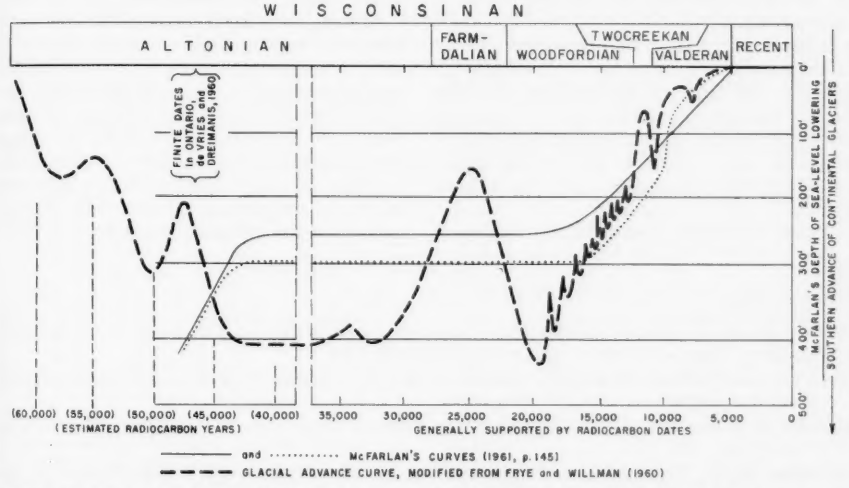

Figure 1. Reproduction of McFarlan's (1961) curves of sea-level lowering in Louisiana on which are superposed curves interpreting continental glacial advance and retreat in the region north of Louisiana

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evidence of this interstadial in Illinois is judged to consist of a minor soil zone (Frye and Willman, 1960; Leonard and Frye, 1960) in the lower part of the Roxana silt, below beds dated 37,000 ± 1500 (W-869) radiocarbon years B.P. The minor fluctuation shown on the curve at an estimated 55,000 radiocarbon years B.P. is inferred from the colluvial zone at the base of the Roxana silt and above the Sangamon soil.

Obviously, McFarlan's scale for sea-level lowering cannot be used for the graphic presentation of glacial advance. As a result, in order to compare these two curves, the glacial curve is arbitrarily adjusted to the space limitations of McFarlan's graph with the bottom of the graph representing maximum southern advance of glaciers and the zero sea-level line representing complete dissipation of continental glaciers. No attempt was made to scale the curve to areas covered by glacial ice. This is admittedly a defect in the comparison of the curves. Glacier thickness and rate and amount of isostatic depression and rebound of the crust are also factors that are not considered. Furthermore, data from central North America only cannot as yet be demonstrated to be a precise replica of world-wide fluctuations of continental glaciers.

In view of these unassessed factors it is not defensible to contend that maximum lowering of sea level should have been at the moment of maximum southern advance (Shelbyville moraine). As indicated by McFarlan, maximum sea-level lowering may very well have been during Altonian time. On the other hand, it seems reasonable that such a major retreat as occurred during Farmdalian time would result in some fluctuation of sea level. The curves appear similar in shape in Figure 1 only for the period of approximately 17,000 B.P. to the present, and here the similarity is more closely related to McFarlan's hinge-line curve than to his mean curve.

Some of McFarlan's datum points fall a considerable distance from his mean curve, and, if his sample points 105, 107, 108, 109, and 49 were connected, the curve might support an interpretation of sea-level oscillation approximately during Farmdalian time.

Although it is freely admitted that no single lobe of a continental glacier can be claimed to serve as a direct index to world-wide fluctuations of sea level, an intimate relationship does exist between Wisconsinan sea-level positions and glacial fluctuations. It is to be hoped that those working in areas of marine sedimentation, as well as in the glaciated regions, will present additional comparisons between well-dated sequences in the two environments.

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Proceedings of the Society

75th Anniversary Volume

The 75th Anniversary of the Society in 1963 is to be marked by the publication of a special volume, which is scheduled for distribution at the time of the Annual Meeting in New York (November 17–20). The planning of the volume is being handled by the 75th Anniversary Committee, which has selected the general subject, outlined the contents, gathered information on authors for the different sections, and established a schedule for the preparation of the manuscripts and the printed book.

The volume will contain a series of short essays, which will explore briefly the philosophy of geology, the methodology and symbolism of geology, and contributions of geology to general knowledge. A selected bibliography of the philosophy of geology will also be included.

The Committee will welcome comments and suggestions on the plan and specifically requests assistance in gathering material for inclusion in the bibliography. In connection with this request, it is noted that much of the pertinent literature may be overlooked because it has appeared in nongeological publications. All communications should be addressed to the Chairman of the 75th Anniversary Committee, Dr. Claude C. Albritton, Department of Geology, Southern Methodist University, Dallas, Texas.

Proceedings of the 57th Annual Meeting of the Cordilleran Section

San Diego State College, San Diego, Calif., March 27–31, 1961

V. L. VanderHoof, *Secretary*

BUSINESS MEETING

The business meeting was held Wednesday noon, March 29; Vice-chairman Ewart Baldwin presided.

The minutes of the 56th Annual Meeting, held at the University of British Columbia, Vancouver, B.C., were approved as read.

The chair then introduced the following persons:

Henry Aldrich, Secretary Emeritus of the Society
Mason Hill, President-elect of the American Association of Petroleum Geologists
Ian Campbell, President of the American Geological Institute
Elmer Fred Davis and William Stephen Webster Kew, both pioneers in the geologic study of California and in the petroleum industry.

Frederick Betz, Jr., the Secretary of the Society, attended the meeting but was forced to return to the home office before the business meeting.

The chair next called for the report of the Nominating Committee, consisting of Charles Chesterman, Chairman, and Messrs. H. V. Warren and George Tunell. They offered the following slate:

Chairman: James Noble
Vice-chairman: Edgar Bailey
Secretary: Richard Fisher

As there were no other nominations, these gentlemen were declared elected by unanimous assent.

Chairman Baldwin then announced that the Section would consider invitations for the next meeting place. Thomas Clements of the University of Southern California renewed his offer of 1960 to meet at his seat of learning, and the Section quickly accepted the offer. The Section will meet there (Los Angeles) in the spring of 1962 at a time to be determined, perhaps Easter recess.

Charles Gilbert then invited the Section to meet at Berkeley in 1963, at which time the new geology building at the University of California will be in use. The Section noted this and will act on it at a later date.

Ian Campbell spoke briefly on A.G.I. affairs and stressed the need for continued support, financially and morally, of this very worth-while clearing house for all branches of geology.

Peter Misch moved that a vote of thanks be given for "the many years of faithful service" of the retiring Section Secretary, V. L. VanderHoof. Carried.

Ben Page moved for a vote of thanks to San Diego State College and its Geology staff for the very successful handling of the 1961 meeting. Carried.

Ian Campbell moved that the Section go on record as having wished Charles Campbell speedy recovery from his current illness. Carried.

The chair then entertained a motion for adjournment, which was made and carried.

ANNUAL DINNER

This traditional affair took place Tuesday evening, March 28, at the El Cortez Hotel. As is customary, it was arranged by the local student geology group—in this case the TECTON-DELVERS.

The speaker of the evening was Prof. Donald B. McIntyre of Pomona College, and he gave a stirring address on James Hutton and the philosophy of Geology. Since a Scot was talking about a Scot, an ancillary feature of the event was the presence of four bagpipers who played loudly and clearly. Not to be outdone, the speaker took a turn with the pipes and seemed quite competent with this difficult pneumatic nightmare.

ATTENDANCE

There were approximately 550 registrants.

SCIENTIFIC SESSIONS

1:30 MONDAY AFTERNOON

Seismology

Room 100, Humanities and Social Science Building

Cochairmen: James T. Wilson and Charles Richter

1. Ari Ben-Menahem: *Deriving fault length and rupture velocity from seismograms: Theory*
2. Frank Press*, Ari Ben-Menahem, and Nafi Toksoz: *Deriving fault length and rupture velocity from seismograms: Experiment*
3. William Stauder, S.J.: *S waves of some Kamchatka earthquakes*
4. Augustine S. Furumoto: *Use of* ScS *wave data for focal-mechanism determinations*
5. Alan S. Ryall: *Direction of faulting in the Hebgen Lake, Montana, earthquake of August 18, 1959*
6. Leonard E. Alsop*, George H. Sutton, and Maurice Ewing: *Free vibrations of the earth observed on strain and pendulum seismographs at Ogdensburg, New Jersey, and Palisades, New York*
7. Stewart W. Smith: *Further study of the earth's free vibrations excited by the Chilean earthquake*

7:30 MONDAY EVENING

Chilean Earthquake Symposium

Music Auditorium

Cochairmen: George Housner and Perry Byerly

1. Pierre St. Amand: *Field observations of the Chilean earthquakes of 1960*
2. Karl V. Steinbrugge: *Chilean earthquakes of May 1960: General engineering view*
3. C. Martin Duke* and David J. Leeds: *Soil, foundation, and earth-structure behavior in the Chilean earthquakes of May 1960*
4. William K. Cloud: *Vibration observations on structures in Chile*
5. Ray W. Clough: *Effect of diagonal braces on the earthquake performance of steel-frame buildings*
6. Doak C. Cox: *Effects of the May 1960 tsunami in Hawaii and other Polynesian islands*

*Speaker

9:00 TUESDAY MORNING

General Session

Little Theater

Chairman: BAYLOR BROOKS

1. Address of Welcome: MALCOLM A. LOVE, President of San Diego State College
2. Introduction of officers of the participating societies
3. General announcements

9:45 TUESDAY MORNING

Sedimentation

Room 210, Business Administration Building

Cochairmen: K. O. EMERY AND M. N. BRAMLETTE

1. EDWARD S. DAVIDSON: *Origin of sediments in the intermontane Safford Basin, Graham County, Arizona* (15 minutes)
2. B. W. CARSS: *Ely Springs dolomite, Arrow Canyon quadrangle, Nevada* (15 minutes)
3. R. L. LANGENHEIM, JR.*, AND H. R. PESTANA: *Upper Ordovician Ely Springs dolomite in eastern Nevada* (15 minutes)
4. K. O. EMERY* AND JOBST HULSEMANN: *Santa Barbara Basin, a semi-euxinic sedimentary environment* (15 minutes)
5. ROBERT J. STANTON, JR.: *Basal sediments of the upper Miocene Castaic formation, Los Angeles County, California* (20 minutes)

9:45 TUESDAY MORNING

Geomorphology

Room 101, Life Sciences Building

Cochairmen: E. L. HAMILTON AND D. M. HOPKINS

1. WILLIAM B. BULL: *Effect of stream-channel shape on alluvial-fan deposition in western Fresno County, California* (20 minutes)
2. THOMAS CLEMENTS: *Pleistocene history of Lake Chapala, Jalisco, Mexico* (20 minutes)
3. MARION T. MILLETT: *Patterns of glacial behavior in coastal Alaska* (20 minutes)
4. FRANCIS P. SHEPARD: *Submarine canyons of the San Diego–La Jolla area* (15 minutes)
5. DAVID M. HOPKINS: *Asymmetrical valleys in Central Alaska* (20 minutes)
6. EARL M. P. LOVEJOY: *Mio-Pliocene Lake Nevada* (15 minutes)

9:45 TUESDAY MORNING

Engineering Geology

Music Auditorium

Cochairmen: E. C. MARLIAVE AND B. E. THOMAS

1. ROBERT G. THOMAS: *Geology of the coastal plain of Los Angeles County as applied to quantitative ground-water studies* (15 minutes)

*Speaker

2. JOHN T. SCHELIGA: *Application of geology to the selection of waste-disposal sites* (15 minutes)
3. PAUL H. JONES: *Application of bore-hole geophysics in hydrologic investigations in a volcanic terrane* (15 minutes)
4. C. L. FAIR: *Geohydrology of the Gila River inner valley near Safford, Arizona* (15 minutes)
5. JAY H. LEHR: *Empirical studies of laminar flow in porous consolidated media* (15 minutes)
6. RICHARD J. PROCTOR: *Engineering geology of Lake Mathews enlargement* (15 minutes)
7. DAVID G. MOORE: *Application of acoustic-reflection techniques to offshore engineering geology* (15 minutes)
8. E. C. MARLIAVE: *Engineering-geology problems in Tunisia* (10 minutes)

8:30 TUESDAY MORNING

Seismology

Room 100, Humanities and Social Science Building

Cochairmen: DON TOCHER AND WILLIAM STAUDER, S.J.

1. WILLIAM R. PERRET* AND BYRON F. MURPHEY: *Transverse surface motion in the immediate vicinity of explosions*
2. D. S. CARDER* AND W. V. MICKEY: *Ground effects from underground explosions*
3. JOSEPH W. BERG, JR.*, AND KENNETH L. COOK: *Energies, magnitudes, and amplitudes of seismic waves from quarry blasts at Promontory and Lakeside, Utah*
4. DORRIS M. HANKINS* AND PAUL R. KINTZINGER: *Seismic data from cratering chemical explosions in Nevada*
5. D. E. WILLIS* AND JAMES T. WILSON: *Effect of decoupling on spectra of seismic waves*
6. JOHN H. HEALY: *Geophysical studies of basin structures along the eastern Sierra Nevada*
7. GEORGE G. SHOR, JR.: *Crustal structure in the San Diego, California, area*

1:30 TUESDAY AFTERNOON

Seismology

Room 100, Humanities and Social Science Building

Cochairmen: DEAN CARDER AND GEORGE SHOR, JR.

1. ROBERT L. KOVACH* AND FRANK PRESS: *Surface-wave dispersion and crustal structure in Antarctica and the surrounding oceans*
2. JAMES DORMAN* AND MAURICE EWING: *Numerical solution of the inversion problem for earthquake surface-wave dispersion*
3. ROBERT A. PHINNEY: *Leaking surface waves on a solid halfspace*
4. JOHN T. KUO* AND JOHN E. NAFE: *Period equation of Rayleigh waves in a layer overlying a halfspace with a sinusoidal interface*
5. JACK OLIVER* AND JAMES DORMAN: *Nature of oceanic seismic surface waves with predominant periods of 6 to 8 seconds*
6. ARMANDO CISTERNAS: *Crustal structure of the Andes from Rayleigh-wave dispersion*
7. C. B. ARCHAMBEAU: *Earth structure from the inversion of surface-wave dispersion data*

*Speaker

1:30 TUESDAY AFTERNOON

Engineering Geology

Music Auditorium

Cochairmen: R. T. Bean and Baylor Brooks

1. Douglas R. Brown: *Geology in the Department of The Los Angeles County Engineer* (15 minutes)
2. J. E. Slosson: *Research in erosion control for hillside housing development* (15 minutes)
3. R. O. Castle: *Problems in engineering geology in the northwest Los Angeles Basin* (15 minutes)
4. Robert Stone: *Geologic and engineering significance of changes in elevation revealed by precise leveling, Los Angeles area, California* (20 minutes)
5. A. L. O'Neill* and R. C. Richter: *Slope failures in foliated rocks of Calaveras group (Carboniferous)* (20 minutes)
6. Carl G. Bock and C. William Schmidt*: *Geology of the Fort Rosecrans landslide, Point Loma, California* (20 minutes)
7. Eugene D. Michael: *Landslides at the Kolob coal field, southwestern Utah* (15 minutes)
8. Harold B. Goldman: *Conglomeratic formations as source of concrete aggregate in California* (15 minutes)
9. James R. Townsend: *Engineering properties of granodiorite from Bernasconi Hills, Riverside County, California* (15 minutes)
10. Parker D. Trask: *Geological problems of constructing an earth dam with aid of nuclear explosions* (15 minutes)

1:30 TUESDAY AFTERNOON

Geochemistry

Room 101, Life Sciences Building

Cochairmen: E. D. Goldberg and F. W. Dickson

1. Carroll J. Halva: *Post lower Miocene potassic basaltic andesites of southeastern Arizona* (20 minutes)
2. L. T. Silver*, C. R. McKinney, and L. A. Wright: *Some Precambrian ages in the Panamint Range, Death Valley, California* (15 minutes)
3. Philip O. Banks* and Leon T. Silver: *Petrological and geochronological observations on the Rubidoux Mountain leucogranites, Riverside County, California* (15 minutes)
4. J. H. Feth*, C. E. Roberson, S. M. Rogers, H. C. Whitehead, and A. S. Van Denburgh: *Westerlies, dust storms and runoff—a geochemical glimpse* (20 minutes)
5. C. W. Blount and F. W. Dickson*: *Solubility of anhydrite in aqueous NaCl solutions at elevated temperatures and pressures* (20 minutes)

1:30 TUESDAY AFTERNOON

Stratigraphy

Room 210, Business Administration Building

Cochairmen: Richard V. Fisher and Richard L. Hay

1. Margaret Fuller Boos: *Petrogenesis of the Precambrian and associated strata west of Denver, Colorado* (15 minutes)

*Speaker

2. Philip H. Heckel* and Anthony Reso: *Silurian and Lower Devonian section in the southwestern part of the Delamar Range, Lincoln County, Nevada* (15 minutes)
3. R. H. Waines: *Devonian sequence in the north portion of the Arrow Canyon Range, Clark County, Nevada* (10 minutes)
4. Anthony Reso: *Silurian stratigraphy in the Pahranagat Range, Lincoln County, Nevada* (15 minutes)
5. Jerry Harbour: *Stratigraphy and sedimentology of the upper Safford Basin, Arizona* (15 minutes)
6. Medora H. Krieger: *Younger Precambrian "Troy" quartzite and the Cambrian formations in the northern end of the Galiuro Mountains, Arizona* (15 minutes)
7. O. E. Bowen* and W. E. Ver Planck: *Stratigraphy of the Oro Grande series in the type locality near Victorville, California* (15 minutes)
8. G. Davidson Woodard: *Stratigraphic succession of the west Colorado Desert, San Diego and Imperial counties, southern California* (15 minutes)
9. J. W. Durham and E. C. Allison*: *Stratigraphic position of the Fish Creek gypsum at Split Mountain gorge, Imperial County, California* (15 minutes)
10. Philip A. Lydon: *Sources of the Tuscan formation in northern California* (20 minutes)
11. Richard L. Hay: *Diagenetic facies in the John Day formation of Oregon* (20 minutes)
12. Gerald R. Licari*, Joan P. Licari, and J. Wyatt Durham: *Geology and amber deposits of the Simojovel area, Chiapas, Mexico* (15 minutes)

8:30 WEDNESDAY MORNING

Mineralogy and Petrology

Room 101, Life Sciences Building

Cochairmen: Celeste Engel and Robert L. Rose

1. René L. Engel: *Granitization and contact metamorphism in the Kern River basin, California* (20 minutes)
2. Robert R. Compton: *Peridotite-gabbro-granodiorite pluton in the Sierra Nevada, California* (20 minutes)
3. Myron G. Best: *Geology of Jurassic (?) rocks in western Mariposa County, southern Sierran foothills, California* (20 minutes)
4. Alden A. Loomis: *Contact metamorphism of the Mt. Tallac roof remnant, Fallen Leaf Lake quadrangle, California* (20 minutes)
5. Robert L. Rose: *"Franciscan" metamorphic rocks near Novato, Marin County, California* (10 minutes)
6. R. G. Coleman and D. E. Lee: *Metamorphic aragonite in the glaucophane schists of Cazadero, California* (15 minutes)
7. Edgar H. Bailey: *Metamorphic facies of the Franciscan formation of California and their geologic significance* (15 minutes)
8. A. Volborth: *Rapakivi granites from Gold Butte, Clark County, Nevada* (20 minutes)
9. Peter Misch: *New criteria for synkinematic growth of metamorphic minerals* (15 minutes)

8:30 WEDNESDAY MORNING

Paleontology

Room 210, Business Administration Building

Cochairmen: York T. Mandra and J. Keith Rigby

1. J. Keith Rigby: *Canadian and Chazyan receptaculitids from Utah and Nevada* (10 minutes)

*Speaker

2. York T. Mandra: *Silicoflagellates: current problems* (20 minutes)
3. W. H. Easton: *New subclass of Paleozoic corals* (15 minutes)
4. Harold J. Bissell: *Fusulinid range zones in Cordilleran area* (20 minutes)
5. Edwin C. Allison, J. Wyatt Durham*, and Victor A. Zullo: *Cold-Water Late Cenozoic faunas of northern California and Oregon* (10 minutes)
6. Augustus K. Armstrong: *Coral zones of the Mississippian Escabrosa limestone of southeastern Arizona and southwestern New Mexico* (15 minutes)
7. James Robert Dodd: *Shell mineralogy and structure as evidence for two subspecies of* Mytilus edulis *on the Pacific Coast of North America* (10 minutes)
8. C. A. Nelson: *Age of the Johnnie formation* (15 minutes)
9. W. P. Kleweno, Jr., and R. M. Jeffords: *Devonian rocks in the Suplee area of central Oregon* (15 minutes)
10. J. R. Macdonald: *Early Miocene mammalian faunas from Wounded Knee, South Dakota* (20 minutes)
11. Theodore Downs* and G. Davidson Woodard: *Middle Pleistocene extension of the Gulf of California into the Imperial Valley* (15 minutes)

8:30 WEDNESDAY MORNING

Areal and Structural Geology

Music Auditorium

Cochairmen: Clarence Allen and Richard P. Phillips

1. Warren Hamilton: *Strike-slip control of tectonics of coastal California* (15 minutes)
2. Bert L. Conrey: *Evidence of Early Pliocene diastrophic activity in the Los Angeles Basin, California* (20 minutes)
3. Dale C. Krause: *Faulting in the continental borderland relative to the Agua Blanca fault* (15 minutes)
4. Blakemore E. Thomas: *Fold structures related to Basin and Range topography near Blythe, California* (15 minutes)
5. Robert L. Wilson: *Geology of the Cushenbury quarry area, San Bernardino County, California: Preliminary report* (15 minutes)
6. Robert L. Christiansen: *Deformation and metamorphism in the El Paso Mountains, Mojave Desert, California* (20 minutes)
7. Peter Lipman: *Isoclinal folding and metamorphism in the eastern Trinity Alps, northern California* (20 minutes)
8. Gregory A. Davis: *"Stratigraphy" and structure of metamorphic rocks in the southwestern quarter of Coffee Creek quadrangle, Trinity Alps, Klamath Mountains, California* (20 minutes)

8:30 WEDNESDAY MORNING

Seismology

Room 100, Humanities and Social Science Building

Cochairmen: Frank Press and John DeNoyer

1. Pierre St. Amand: *Geologic and tectonic framework of the Chilean earthquakes of May 1960*
2. Admiral William M. Gibson: *Geodimeter measurements across the San Andreas fault, California*

*Speaker

3. W. C. Meecham, John DeNoyer, D. E. Willis, and James T. Wilson*: *Azimuthal asymmetry of a point source in a two-dimensional low-velocity medium*
4. Freeman Gilbert and Stanley J. Laster*: *Excitation and propagation of pulses on an interface*
5. Otto Nuttli* and John D. Whitmore: *Observations of* S *wave motion in the distance interval of 50° to 75°*
6. Gordon E. Frantti,* D. E. Willis, and James T. Wilson: *Spectrum of seismic noise*
7. George Backus* and Freeman Gilbert: *Rotational splitting of the free oscillations of the earth*

1:30 WEDNESDAY AFTERNOON

Seismology

Room 100, Humanities and Social Science Building

Cochairmen: Jack Oliver and Hugo Benioff

1. Iya Abubaker* and John H. Healy: *A seismic model study of the effect of basin structures on the phase velocity of Rayleigh waves*
2. James N. Brune,* John E. Nafe, and Leonard E. Alsop: *Polar-phase shift of surface waves on a sphere*
3. Frank Press, David Harkrider,* and C. A. Seafeldt: *Computation of dispersion of surface waves on digital computers*
4. Freeman Gilbert* and Stanley J. Laster: *Experimental investigation of* PL *modes in a single layer*
5. Robert Gunst and R. J. Brazee*: *Coast and Geodetic Survey method of earthquake hypocenter location by electronic computer*
6. Louis W. Erath: *Electronic seismometer*
7. E. A. Flinn,* C. F. Romney, A. M. Rugg, and J. R. Woolson: *Analytical results from low-frequency seismic measurements in a deep well*
8. George H. Sutton* and Paul W. Pomeroy: *Long-period magnetic-tape seismograph system: First results*

1:30 WEDNESDAY AFTERNOON

Areal and Structural Geology

Music Auditorium

Cochairmen: Ben M. Page and Ellis Roberts

1. Robert S. Dietz: *Astroblemes: Ancient meteorite impact structures* (20 minutes)
2. Ben M. Page: *Gravity tectonics southwest of Parma, Northern Apennines, Italy* (15 minutes)
3. Robert S. Yeats: *Structure of the Cascade Range east of Seattle, Washington* (20 minutes)
4. Ross C. Ellis: *Tertiary deformation in the Central Washington Cascade Range* (10 minutes)
5. William A. Oesterling: *Northwest thrusting in northeast Nevada: Wood Hills thrust* (20 minutes)
6. Lehi F. Hintze: *New geologic map of Western Utah* (15 minutes)
7. R. J. Roberts* and E. W. Tooker: *Structural geology of the northern Oquirrh Range, Utah* (15 minutes)

*Speaker

8. E. W. Tooker* and R. J. Roberts: *Stratigraphy of the northern Oquirrh Range, Utah* (15 minutes)
9. Philip B. King: *Neo-tectonic map of the U.S.S.R.* (20 minutes)

1:30 WEDNESDAY AFTERNOON

Mineralogy and Petrology

Room 101, Life Sciences Building

Cochairmen: D. O. Emerson and John Christie

1. Neville L. Carter: *Petrology of the Venås granite and the surrounding rocks, east Telemark, Norway* (20 minutes)
2. D. F. Davidson* and H. W. Lakin: *Metalliferous shales in the United States* (15 minutes)
3. William R. Dickinson: *Jurassic andesitic province along the Pacific margin of North America* (20 minutes)
4. Lewis S. Lohr and David B. Slemmons*: *Structural state of plagioclase in hydrothermally altered rocks from Pyramid Butte, Washoe County, Nevada* (15 minutes)
5. Vincent P. Gianella: *Fused tuff in the Hays Canyon Range, Nevada* (15 minutes)
6. D. O. Emerson: *Sodic plagioclase of intermediate temperature state in the "Nomlaki tuff" near Winters, California* (10 minutes)
7. Donald B. Tatlock* and James O. Berkland: *Leona "soda" rhyolite, Alameda County, California—primary or metasomatic?* (15 minutes)
8. Bates McKee: *Widespread distribution of aragonite in the Pacheco Pass area, California* (15 minutes)
9. Paul H. Reitan: *"Hot spots"—of significance in recrystallization?* (15 minutes)
10. Alexander K. Baird: *Rapid and precise quantitative analysis of major elements in granitic rocks by X-ray fluorescence spectrography* (15 minutes)
11. Lon S. McGirk, Jr.: *Heavy-liquid diluents and technique* (5 minutes)
12. John B. Adams: *Transition from Skagit gneiss to directionless Black Peak quartz diorite, north of Lake Chelan, Northern Cascades, Washington* (15 minutes)

*Speaker

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REVIEWS IN GEOLOGY

The following Review Articles, published in this issue and in the June 1960 Bulletin, are available as separates. The price is 75 cents per copy.

REVIEW OF PALEOMAGNETISM. Allan Cox and Richard R. Doell

RIVER MEANDERS. Luna B. Leopold and M. Gordon Wolman

ASH FLOWS. Robert L. Smith

PALEOTECTONIC EVOLUTION OF THE CENTRAL AND WESTERN ALPS. Rudolf Trümpy

FACTORS INFLUENCING SUPPLY OF MAJOR IONS TO INLAND WATERS, WITH SPECIAL REFERENCE TO THE ATMOSPHERE. Eville Gorham

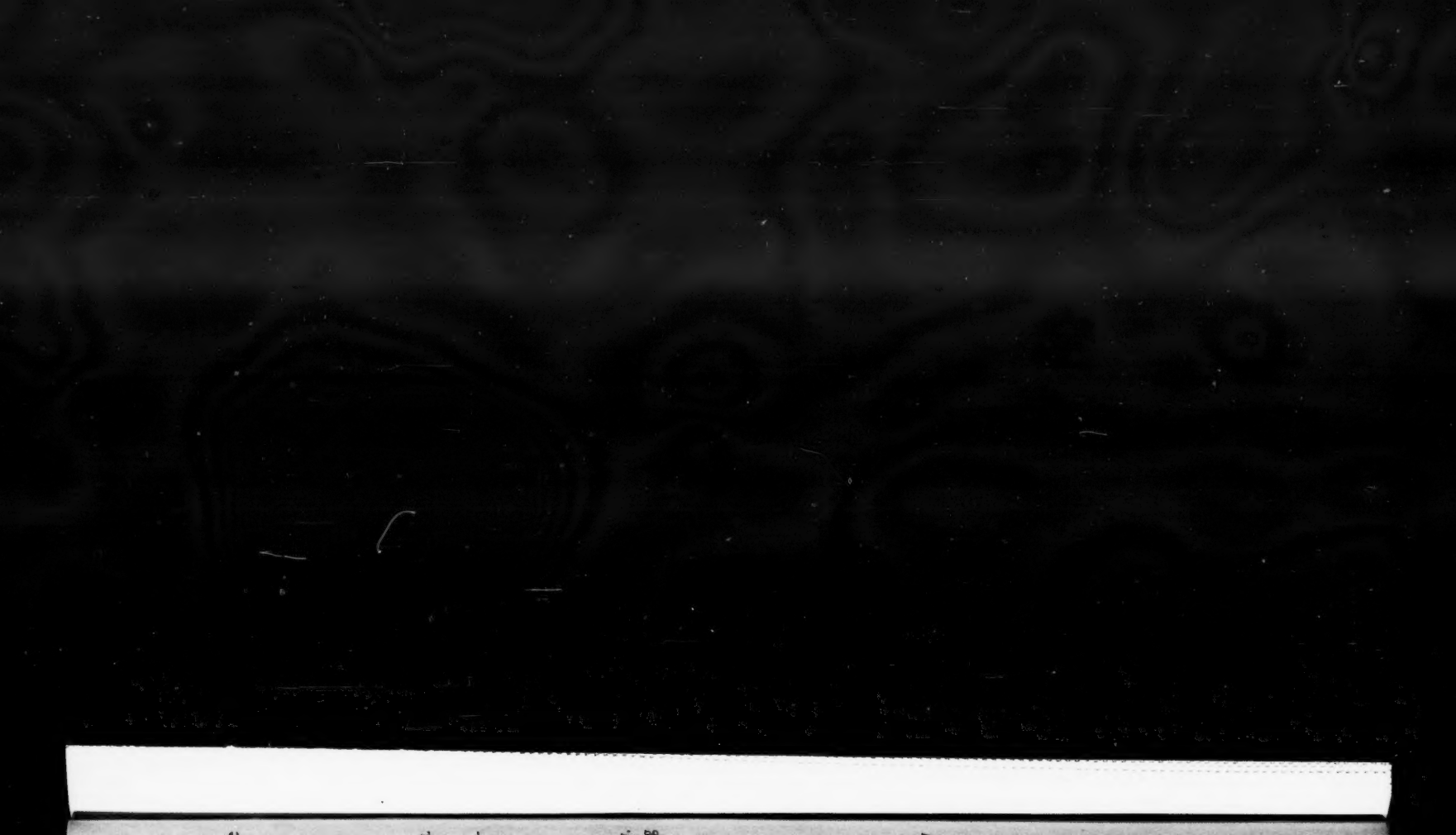
SYMMETRY CONCEPTS IN THE STRUCTURAL ANALYSIS OF DEFORMED ROCKS. M. S. Paterson and L. E. Weiss

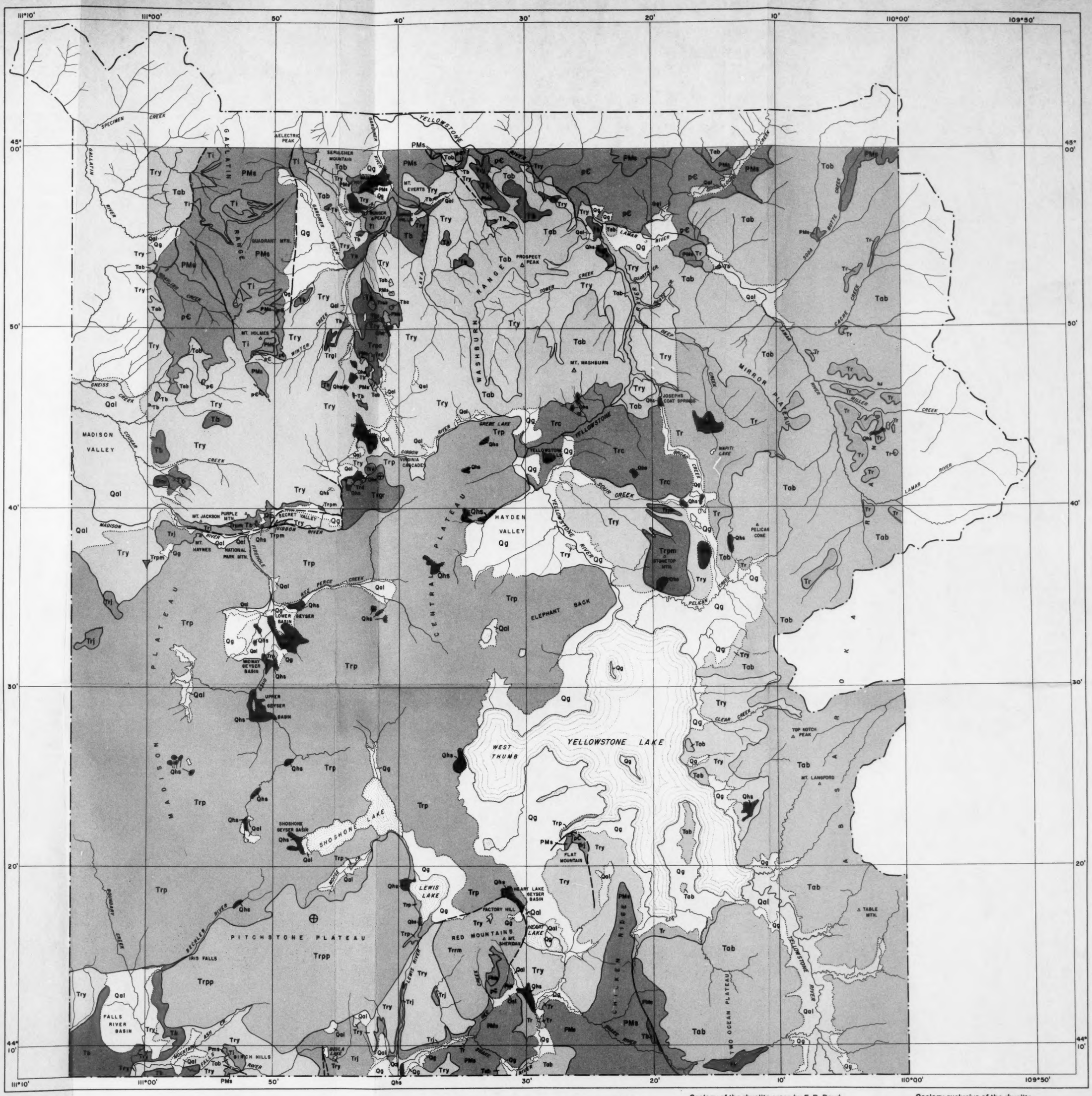
EXPLORATION GEOPHYSICS: A REVIEW. Frederick E. Romberg

LATE PLEISTOCENE OF EUROPE: A REVIEW. H. E. Wright, Jr.

Additional Review Articles published in earlier volumes of the Bulletin are listed below. They are not available as separates, but those in stock may be obtained by purchasing the issue of the Bulletin in which they appeared.

FOSSILS IN METAMORPHIC ROCKS: A REVIEW. W. H. Bucher NONRADIOGENIC ISOTOPES IN GEOLOGY: A REVIEW. Earl Ingerson METAMORPHISM OF BASALTIC ROCKS: A REVIEW. Arie Poldervaart	V. 64, no. 3; $2.50 per copy
PERMAFROST: A REVIEW. R. F. Black GLACIER FLOW: A REVIEW. R. P. Sharp	V. 65, no. 9 (Out of stock)
CARBONATITES: A REVIEW. W. T. Pecora	V. 67, no. 11; $2.50 per copy
GRANITE EMPLACEMENT WITH SPECIAL REFERENCE TO NORTH AMERICA. A. F. Buddington ION EXCHANGE IN CLAYS AND OTHER MINERALS. Dorothy Carroll GEOLOGIC SIGNIFICANCE OF ACCUMULATOR PLANTS IN ROCK WEATHERING. T. S. Lovering	V. 70, no. 6; $2.50 per copy





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Geology of the rhyolite areas by F. R. Boyd, R. Schmalz, R. Lundin, P. Calkin

Geology exclusive of the rhyolite after U. S. G. S. Folio no 30

QUATERNARY

- Qal — Alluvium
- Qhs — Hot spring deposits
- Qg — Glacial and glacial lake deposits

PLIOCENE

- Trcc — Obsidian Cliff flow; rhyolite; Plateau flows
- Trcs — Crystal Springs flow; rhyolite; Plateau flows
- Trcc — Cougar Creek flow; rhyolite; Plateau flows
- Trc — Canyon flow; rhyolite; Plateau flows

PLIOCENE

- Trpp — Pitchstone Plateau flow; rhyolite; Plateau flows
- Trgr — Gibbon River flow; rhyolite; Plateau flows
- Trp — Plateau flows; undifferentiated; rhyolite
- ⊕ — Vent for Plateau flow
- Trg — Grizzly Lake rhyolite–basalt complex
- Tbc — Basaltic cinder cone

PLIOCENE

- Tb — Basalt; both pre and post Yellowstone tuff
- Try — Yellowstone tuff; welded rhyolite tuff
- Trpm — Purple Mountain pumice breccia; rhyolite
- Trd — Rhyolite dome
- Trj — Jackson flows; undifferentiated; rhyolite
- Trrm — Red Mountains rhyolite; flows and flow breccias; undifferentiated
- Tr — Rhyolite; undifferentiated; pre Plateau flows

EARLY TERTIARY

- Tab — Andesite breccia and basalt
- Ti — Intrusive rocks

PALEOZOIC MESOZOIC

- PMs — Sedimentary rocks

PRE-CAMBRIAN

- pC — Gneiss and schist

— — Concealed fault

GEOLOGIC MAP OF YELLOWSTONE NATIONAL PARK, WYOMING

BOYD, PLATE 1
Geological Society of America Bulletin, volume 72

EXPLANATION

QUATERNARY

- Qal — Alluvium
- Hot spring deposits
- Qg — Glacial deposits

PLIOCENE

- Trgr — Gibbon River flow; rhyolite ; Plateau flows
- Trvd — Vent dome of Gibbon River flow
- Trcc — Cougar Creek flow ; rhyolite ; Plateau flows
- Troc — Obsidian Cliff flow; rhyolite ; Plateau flows
- Trcs — Crystal Springs flow ; rhyolite ; Plateau flows
- Trp — Plateau flows ; rhyolite ; undifferentiated
- Trgl — Grizzly Lake rhyolite—basalt complex
- Tbc — Basaltic cinder cone

PLIOCENE

- Basalt
- Try — Yellowstone tuff ; welded rhyolite tuff
- Trpm — Purple Mountain pumice breccia ; rhyolite
- Rhyolite dome
- Trj — Jackson flows ; rhyolite ; undifferentiated

EARLY TERTIARY

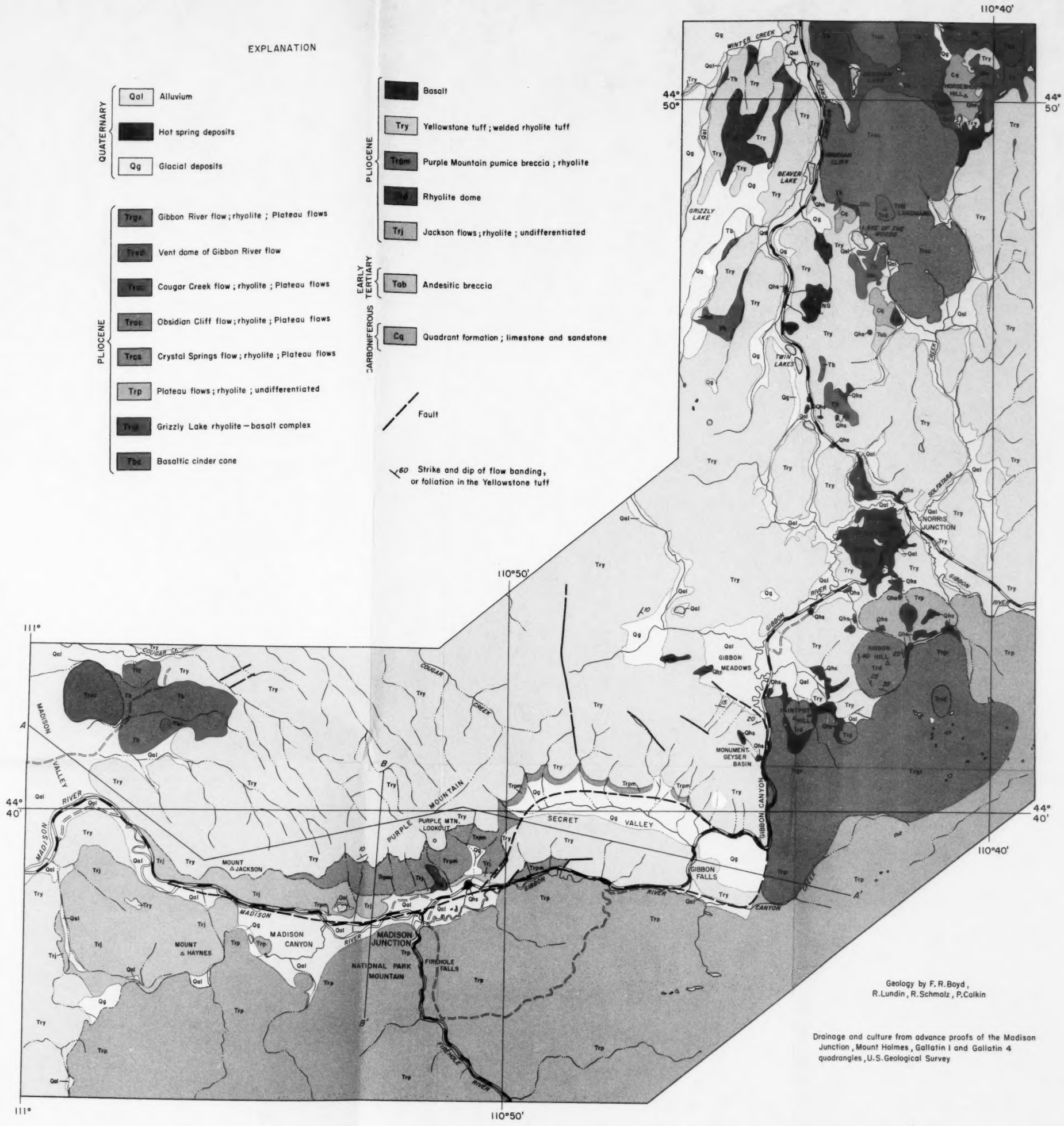
- Tab — Andesitic breccia

CARBONIFEROUS

- Cq — Quadrant formation ; limestone and sandstone

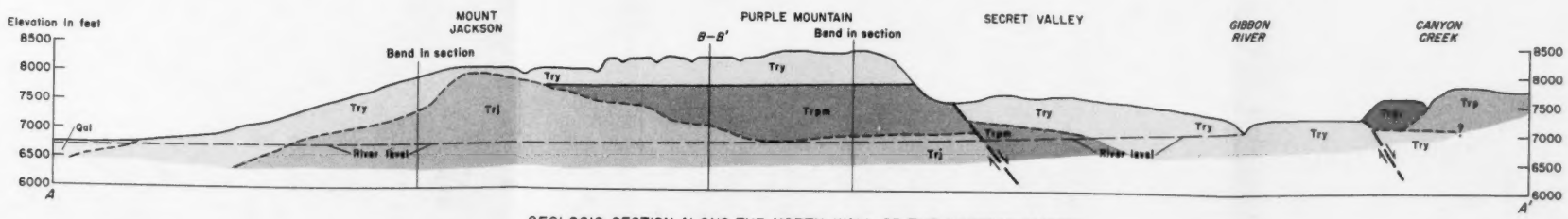
Fault

60 Strike and dip of flow banding, or foliation in the Yellowstone tuff

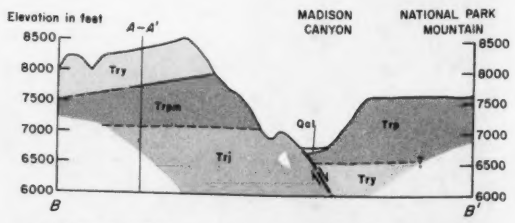


Geology by F. R. Boyd, R. Lundin, R. Schmalz, P. Calkin

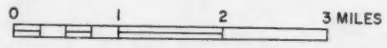
Drainage and culture from advance proofs of the Madison Junction, Mount Holmes, Gallatin I and Gallatin 4 quadrangles, U.S. Geological Survey



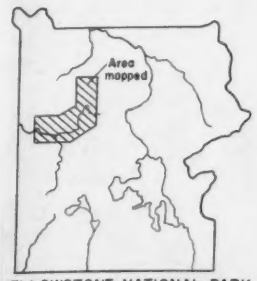
GEOLOGIC SECTION ALONG THE NORTH WALL OF THE MADISON CANYON



GEOLOGIC CROSS SECTION OF THE MADISON CANYON



SCALE FOR MAP AND HORIZONTAL SCALE FOR SECTIONS



YELLOWSTONE NATIONAL PARK

GEOLOGIC MAP OF THE MADISON CANYON, NORRIS GEYSER BASIN, AND OBSIDIAN CLIFF AREAS, YELLOWSTONE NATIONAL PARK, WYOMING

BOYD, PLATE 6
Geological Society of America Bulletin, volume 72

TABLE 1.—ANALYSES OF ROCKS FROM UMNAK AND BOGOSLOF ISLANDS, ALEUTIAN ISLANDS, ALASKA

Abundance of minor elements determined spectrographically.

	Northeastern Umnak																				Southwestern Umnak									
	1	1A	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29
Major Oxides (Weight per cent)																														
	SMB[a] RLH LMS	EJT	WWH	SMB RLH LMS	MKC	WWB LMS	FMB WWH	FMB WWH	SMB RLH LMS	FMB WWH	WWH	WWH	WWB LMS	FMB WWH	FMB	FMB WWH	FMB WWH	SMB RLH LMS	SMB RLH LMS	FMB	MKC	WWB LMS	WWB LMS	WWB LMS	WWB LMS	WWB LMS	SMB RLH LMS	SMB RLH LMS	SMB RLH LMS	SMB RLH LMS	
SiO_2	45.7	48.07	48.43	48.7	48.97	50.1	50.52	50.60	51.4	52.23	52.43	52.58	53.1	54.35	54.41	54.80	57.40	63.5	65.6	72.36	72.77	54.3	56.0	57.4	57.5	60.6	61.3	61.4	64.4	65.6
Al_2O_3	12.9	15.61	18.94	14.9	16.27	23.0	19.71	18.06	14.4	17.04	15.27	14.97	18.4	16.38	15.77	14.90	15.49	16.1	14.1	13.75	13.29	18.0	15.6	18.2	17.4	15.4	16.3	16.5	16.0	14.1
Fe_2O_3	3.6	2.34	4.15	4.7	2.83	2.6	1.67	2.99	6.3	3.22	5.96	3.86	4.5	3.01	4.27	2.88	1.97	1.4	2.0	0.59	1.48	1.7	2.1	2.0	2.2	2.7	2.1	2.6	2.1	4.5
FeO	5.6	6.98	6.30	3.4	6.24	5.4	6.82	6.75	1.3	6.60	6.99	8.81	3.2	8.00	6.55	9.00	7.86	5.0	3.8	2.33	1.55	7.8	4.8	5.0	5.2	5.6	6.0	5.4	4.1	2.2
MgO	18.2	12.68	6.55	4.2	9.62	3.6	4.64	5.58	1.5	5.54	4.13	4.00	3.4	3.34	4.02	3.00	2.94	1.2	0.49	0.09	0.08	3.8	8.8	4.0	3.8	2.0	1.6	2.2	1.2	2.0
CaO	9.6	10.76	11.04	7.8	12.86	11.6	11.96	11.46	8.4	10.39	8.70	8.64	9.3	8.08	8.29	7.15	6.74	3.8	2.8	1.53	1.27	7.9	7.7	7.7	6.9	3.5	5.6	5.6	4.2	2.8
Na_2O	1.7	1.86	2.51	2.1	2.00	2.4	2.71[b]	2.73	4.4	2.97	3.50	3.47	2.8	3.66[b]	3.50	3.83	3.96	4.6	4.3	4.64	4.72	3.4	2.4	2.9	3.6	5.4	4.6	4.3	4.5	6.1
K_2O	0.52	0.98	0.52	0.92	0.54	0.44	0.58[b]	0.53	1.1	0.78	1.24	1.22	1.4	1.32[b]	1.26	1.23	1.53	2.6	3.0	4.16	4.06	1.0	1.6	1.4	1.7	2.2	1.7	1.6	2.6	1.5
Ignition[c] H_2O+	..	0.09	0.12	..	0.09	..	0.08[b]	0.17	..	0.12[b]	0.20	0.20	1.14	0.08[b]	0.09	0.18[b]	0.20[b]	..	..	0.27	0.10	..	..	..	..	1.07	..	..	..	..
Ignition[c] H_2O--	−0.16[d]	0.07	0.11	10.1	0.04	−0.13[d]	0.01[b]	0.05	6.3	0.06[b]	0.11	0.07	0.54	0.04[b]	0.02	0.09[b]	0.09[b]	−0.08[d]	3.4	0.03	0.02	−0.20[d]	0.03	0.20	0.24	0.31	−0.51[e]	−0.29[e]	−0.11[d]	1.4
Ignition[c] CO_2	..	0.01	..	..	..	..	..	..	..	..	..	..	0.65	..	..	..	..	..	..	..	..	..	..	..	..	0.07	..	..	..	..	..
TiO_2	0.51	0.73	1.20	1.2	0.72	0.74	0.97	1.10	0.93	1.24	1.48	1.80	1.0	1.25	1.56	2.48	1.54	0.81	0.64	0.21	0.21	1.2	0.66	0.90	1.2	1.0	1.0	1.1	0.82	0.75
P_2O_5	0.10	0.12	0.27	0.22	0.13	0.11	0.19[b]	0.28[b]	2.4	0.25[b]	0.36	0.38	0.30	0.33[b]	0.25	0.65	0.32[b]	0.23	0.16	0.06	0.06	0.32	0.17	0.16	0.4	0.43	0.42	0.18	0.23	0.20
MnO	0.19	0.17	0.17	0.20	0.18	0.13	0.18[b]	0.16[b]	0.23	0.19[b]	0.17	0.21	0.14	0.15[b]	0.21	0.25	0.18[b]	0.22	0.18	0.09	0.10	0.18	0.12	0.10	0.10	0.15	0.20	0.17	0.16	0.11
Total	98.62	99.97	100.31	98.44	100.49	100.12	100.04	100.46	98.66	100.63	100.54	100.21	99.85	99.99	100.20	100.44	100.22	99.46	100.47	100.11	99.71	99.60	99.98	99.96	100.24	100.43	100.82	101.05	100.31	101.26
Powder Sp. Gr.	3.02	..	2.97	2.49	3.0[e]	2.80	2.86	2.96	2.55	2.94	2.89	2.93	2.78	2.88	2.8[e]	2.92	2.83	2.58	2.45	2.4[e]	2.4[e]	2.79	2.88	2.81	2.84	2.76	2.59	2.67	2.59	2.71
Minor Elements[f]																														
Zr		PRB[a] .003			AC .004	ATM .003	KJM .005				KJM .007	KJM .008	ATM .004			KJM .01	KJM .01			KJM .02		ATM .007		ATM .005	AC .007	ATM .008					
Ga		.002			.0006	.0009	.0007				.0007	.001	.001			.001	.0007			.001		.001		.0008	.002	.001					
La		..			..	..	..				..	..	..			..	..			..		..		..	..	..					
Sc		..			.004	..	..				..	..	..			..	..			..		..		..	.002	..					
Y		..			.003	.002	.002				.005	.006	.002			.007	.007			.01		.004		.002	.004	.004					
Yb		..			.0002	..	..				..	..	..			..	..			..		..		..	.0003	..					
B		..			..	..	..				..	..	..			..	..			.006		.002		.002	.002	..					
Be		..			..	..	..				..	..	..			..	..			.0005		..		..	..	..					
Sr		.08			.1	.05	.09				.1	.1	.05			.06	.1			.02		.05		.03	.06	.03					
Ba		.06			.03	.01	.02				.03	.03	.05			.03	.05			.09		.02		.05	.02	.06					
Pb		..			.001	..	..				..	..	..			.001	..			.005		..		..	..	..					
Mo		..			..	..	..				..	..	..			..	..			..		..		..	.0006	..					
Cu		.004			.02	.003	.02				.03	.04	.007			.01	.008			.004		.009		.006	.005	.003					
V		.03			.03	.02	.02				.03	.05	.02			.02	.02			.001		.02		.01	.02	.02					
Cr		.07			.03	.004	.008				.004	.01	.007			..	.003			.0003		.0007		.007	.01	.0003					
Ni		.03			.01	.001	.004				.002	.004	.003			..	.001			..		.001		.004	.003	.001					
Co		.007			.004	.002	.002				.002	.003	.002			.002	.002			..		.002		.002	.003	.001					
Nb		..			..	..	..				..	..	..			..	..			..		..		..	.002	..					
Total		.286			.2368	.0959	.1717				.2107	.252	.146			.141	.2017			.1618		.1167		.1208	.1409	.1283					
C. I. P. W. Norms (Weight Per Cent)																														
Q	..	..	..	12.1	..	2.5	..	1.2	4.6	2.8	4.9	3.7	10.1	4.8	6.7	7.7	8.3	14.5	21.5	25.2	27.0	5.1	6.0	11.3	9.2	9.5	12.6	14.0	16.5	17.2
C	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
or	2.8	2.8	3.1	5.6	2.8	2.8	3.5	3.0	6.7	4.5	7.3	7.2	8.3	7.8	7.5	7.2	8.9	15.6	17.8	24.6	24.5	6.1	9.4	8.3	10.0	12.8	10.0	9.5	15.6	8.9
plag ab	14.2	15.7	21.2	17.8	16.8	20.4	22.9	23.1	37.2	25.1	29.6	29.4	23.6	30.9	29.6	32.5	33.5	38.8	36.1	39.3	39.8	28.8	20.4	24.6	30.4	45.6	38.8	36.2	38.3	51.4
plag an	26.1	32.8	38.9	28.4	34.2	50.3	39.9	35.5	16.1	30.8	22.3	21.7	33.4	24.7	23.6	19.7	20.0	15.6	10.3	4.4	2.8	30.6	27.0	32.3	26.4	11.4	18.9	21.1	15.6	6.7
plag ne	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
di wo	8.6	8.2	5.9	3.6	12.1	2.7	7.6	8.1	4.2	7.9	7.7	7.8	3.0	5.5	6.6	4.7	4.8	1.0	1.2	1.2	1.2	2.9	4.4	2.2	2.1	1.2	2.6	2.4	1.6	2.7
di en	6.6	5.7	3.9	2.9	8.3	1.4	4.0	4.8	3.6	4.8	5.6	3.7	2.4	2.4	3.9	1.9	1.9	0.3	0.3	0.1	0.17	1.3	3.1	1.3	1.2	0.5	0.9	1.2	0.7	2.3
di fs	1.1	1.8	1.6	0.3	2.8	1.2	3.4	2.8	..	2.7	2.6	3.9	0.3	3.1	2.4	2.8	2.8	0.8	0.9	1.2	1.1	1.6	0.9	0.8	0.8	0.7	1.7	1.2	0.9	..
hy en	1.7	8.7	10.6	7.6	6.6	7.6	7.5	9.0	0.1	9.0	5.6	6.2	6.1	5.9	6.1	5.5	5.4	2.7	0.9	0.2	0.03	8.2	18.9	8.7	8.3	4.5	3.1	4.3	2.3	2.7
hy fs	0.3	2.8	4.2	0.5	2.2	5.9	6.4	5.6	..	5.1	3.2	6.6	0.4	7.4	3.9	7.8	7.8	6.2	3.6	2.4	0.22	9.5	5.3	5.3	5.0	5.9	6.2	5.0	3.8	..
ol fo	26.1	12.0	1.3	..	6.3	..	0.1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
ol fa	4.3	4.1	0.5	..	2.4	..	0.1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
mt	5.3	3.5	6.0	6.7	4.2	3.7	2.4	4.3	2.3	4.7	8.7	5.6	6.5	4.4	6.2	4.2	2.9	2.1	3.0	0.9	2.1	2.6	3.0	3.0	3.3	3.9	3.0	3.7	3.0	5.3
il	0.9	1.4	2.3	2.3	1.4	1.4	1.8	2.1	1.7	2.4	2.8	3.4	2.0	2.4	2.9	4.7	2.9	1.5	1.2	0.4	0.4	2.4	1.2	1.5	2.3	2.0	2.0	2.1	1.5	1.4
hm	..	..	..	..	..	..	..	..	10.1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	0.8
ap	0.3	0.3	0.6	0.6	0.3	0.3	0.4	0.7	5.7	0.6	0.8	0.9	0.7	0.8	0.6	1.6	0.8	0.3	0.3	0.1	0.1	0.7	0.3	0.3	1.0	1.0	1.0	0.3	0.5	0.4
cc	..	..	..	..	..	..	..	..	..	..	..	..	1.6	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..
Total	98.3	99.8	100.1	88.4	100.4	100.2	100.0	100.2	92.3	100.4	100.2	100.1	98.4	100.1	100.0	100.3	100.0	99.4	97.1	100.0	99.42	99.8	99.9	99.6	100.0	99.0	100.8	101.0	100.3	99.8
Niggli Values																														
si	83	98	110	150	103	124	122	120	175	129	134	136	148	148	146	155	170	236	286	372	384	148	141	167	169	207	203	200	240	248
al	14	19	25	27	20	33	28	25	29	25	23	23	30	26	25	25	27	35	36	42	42	29	23	31	30	31	32	31½	35	31½
fm	64	54	42	39	46	29½	34	39	24	40½	42	42	32	38	40	41	37½	27	24½	13	13	37	47½	34	35	33½	30	32	26	31½
c	19	23	27	26	29	31	31	29	31	27	24	24	28	24	24	22	21½	15	13	8	7	23	21	24	22	13	20	19½	17	11
alk	3½	4	6	8	5	6½	7	7	17	7½	11	11	10	12	11	12	14	23	26½	37	38	11	8½	11	13	22½	18	17	22	26
k	0.16	0.14	0.12	0.23	0.14	0.11	0.12	0.11	0.14	0.15	0.19	0.19	0.25	0.19	0.19	0.17	0.20	0.27	0.32	0.37	0.37	0.17	0.30	0.24	0.24	0.21	0.20	0.20	0.28	0.14
mg	0.78	0.72	0.53	0.49	0.66	0.45	0.49	0.51	0.32	0.50	0.37	0.36	0.45	0.35	0.40	0.31	0.35	0.25	0.13	0.05	0.05	0.42	0.70	0.51	0.48	0.30	0.26	0.34	0.26	0.36
qz	−31	−18	−14	18	−17	−2	−6	−8	7	−1	−10	−8	8	0	2	7	14	44	80	124	134	4	7	23	17	17	27	32	50	44

[a] Analysts: S. M. Berthold (SMB), W. W. Brannock (WWB), R. L. Hill (RLH), and Leonard M. Shapiro (LMS), Rapid Analysis Laboratory (*See* Shapiro and Brannock 1956), U. S. Geological Survey; W. W. Hommel (WWH), Ledoux and Company; T. M. Chatard (TMC), M. K. Carron (MKC), and E. J. Tomasi (EJT), U. S. Geological Survey; F. M. Byers, Jr., (FMB) at University of Chicago laboratory; Paul R. Barnett (PRB), Arthur Chodos (AC), K. J. Murata (KJM), A. T. Myers (ATM), spectrographers, U. S. Geological Survey

[b] These constituents determined by W. W. Hommel, Ledoux and Company

[c] Includes gain due to oxidation of FeO

[d] Gain due to oxidation of FeO greater than losses of volatile constituents; not included in totals

[e] Inferred from bulk specific gravity

[f] Approximate lower limit of sensitivity (in per cent) for minor elements looked for but not found: La (0.003), Sc (0.0002), Yb (0.0001), B (0.001), Be (0.0001), Pb (0.001), Mo (0.0005), V (0.001), Cr (0.0001), Ni (0.0001), Co (0.0001), Nb (0.002), Sn (0.002), As (0.05), Sb (0.01), Bi (0.001), Ge (0.001), In (0.001), Tl (0.005), and Re (0.005)

1. Picritic basalt flow (45–95) of Ashishik basalt at Cape Idak (lat. 53°31.4′ N., long. 167°47.8′ E.)

1A. Olivine-rich basalt flow (46 ABy 531) of Ashishik basalt near Cape Idak (lat. 53°31.3′ N., long. 167°47.8′ E.)

2. Plagioclase-olivine basalt flow (46 ABy 92) of Mt. Tulik (lat. 53°22.4′ N., long. 167°04.3′ E.)

3. Palagonitized tuff cone (46 ABy 132) of early postcaldera pyroclastic rocks within Okmok Caldera (lat. 53°26.4′ N., long. 168°05.0′ E.), 1 mile northeast of cone D (Pl. 1). Specimen collected from palagonitized bed near summit.

4. Anorthite-augite-olivine basalt flow (46 AFr 28) of Ashishik basalt (lat. 53°27.7′ N., long. 167°55.5′ E.), 3 miles N. 25° W. of Umnak Airfield. Specimen collected from lower middle part of flow.

5. Anorthite basalt flow (46 ABy 609) near Inanudak Bay (lat. 53°17.7′ N., long. 168°21.3′ E.), from sea cliff S. 25° E. of cinder cone

6. Anorthite basalt flow (46 AHp 68) extruded in December 1945 from cone A (Pl. 1) (lat. 53°24.7′ N., long. 168°11.0′ E., 6000 feet northwest of cone)

7. Bytownite-augite-olivine basalt flow (46 ABy 29), a subaerial flow of cone D (Pl. 1) (lat. 53°25.6′ N., long. 168°05.8′ E., south-southeast slope, altitude 1900 feet)

8. Palagonitized scoria bed (45–73) from Ashishik basalt (lat. 53°21.2′ N., long. 168° 17.0′ E.)

9. Augite basalt flow (46 ABy 70) of Ashishik basalt in lower part of Okmok Caldera wall (lat. 53°27.3′ N., long. 168°05.5′ E.)

10. Flow no. 2 (46 ABy 45) of Crater Creek basalt (lat. 53°28.5′ N., long. 168°05.0′ E.) in northwest wall of Crater Creek gorge

11. Flow no. 11 (46 ABy 54) of Crater Creek basalt (lat. 53°28.7′ N., long. 168°04.6′ E.) in northwest wall of Crater Creek gorge

12. Hydrothermally altered plagioclase basalt (47 ABy 44) from volcanic rocks of central Umnak. From cliff just south of Air Force hut near Inanudak Bay (lat. 53°17.1′ N., long. 168°19.8′ E.)

13. Lower welded andesitic agglomerate bed (46 ABy 13) in Okmok volcanics exposed in upper part of north wall (lat. 53°28.3′ N., long. 168°08′ E.) of Okmok caldera. Sample taken to exclude most foreign lithic fragments.

14. Older basaltic andesite flow (46 ABy 6) from cone B (Pl. 1) in Okmok Caldera. Near front of flow (lat. 53°27.6′ N., long. 168°07′ E.)

15. Aphyric andesite pipe (46 ABy 63) of vitreous andesite. Near top of pipe (lat. 53°28.0′ N., long. 168°05.4′ E.) at intersection of southeast wall of Crater Creek gorge and Okmok Caldera

16. Upper welded andesitic agglomerate bed (46 ABy 15) in Okmok volcanics exposed in upper part of north wall (lat. 53°28.3′ N., long. 168°08′ E.) of Okmok Caldera. Sample taken to exclude most xenolithic fragments.

17. Vitreous latite fragment (46 AFr 2) in Okmok volcanics. From bank of Crater Creek (lat. 53°32.6′ N., long. 167°59.2′ E.) near Cape Tanak

18. Rhyodacite pumice lapilli bed (47 ABy 26) near base of Okmok volcanics. Sea cliff (lat. 53°33.5′ N., long. 168°04.9′ E.) 0.6 mile S. 10° E. of Ashishik Point

19. Rhyolite obsidian band (46 ABy 30a) in rhyolite from northeastern Umnak (lat. 53°29.5′ N., long. 168°10.3′ E.)

20. Felsitic rhyolite facies (48 Um 11) in east part of rhyolite from northeastern Umnak (lat. 53°29.4′ N., long. 168°09′ E.)

21. Basaltic andesite flow (47 ABy 73) of Mount Vsevidof from Twinlava Point (lat. 53°11.1′ N., long. 168°47.7′ E.)

22. Quartz-bearing olivine andesite flow (47 ABy 10). North-facing cliff (lat. 53°13.6′ N., long. 168°21.9′ E.), south of Inanudak Bay

23. Hypersthene-bearing labradorite andesite flow (47 ABy 9) of Mount Recheschnoi (lat. 53°13.8′ N., long. 168°22.0′ E.), south of Inanudak Bay

24. Quartz diorite (47 ABy 41) facies of plutonic rocks (lat. 53°16.1′ N., long. 168°34.9′ E.), 1¾ miles S. 70° E. of Cape Ilmalianuk

25. Albitized diorite (47 ABy 34) in albitized sedimentary and igneous complex. From south shore line (lat. 52°51.8′ N., long. 169°00.1′ E.) 1½ miles east of Idaliuk Point

26. Andesitic glass clot (47 ABy 72) in dark scoria bed of Mount Vsevidof cone. Specimen collected at an altitude of 4000 feet on north slope of cone (lat. 53°07.7′ N., long. 168°41.3′ E.)

27. Hypersthene-bearing andesite flow (47 ABy 57) of Mount Recheschnoi, south slope (lat. 53°05.3′ N., long. 168°28.5′ E.)

28. Historic(?) latite flow (47 ABy 36) of Mount Vsevidof. Front of flow (lat. 53°07.7′ N., long. 168°46.6′ E.) on west slope of Mount Vsevidof

29. Quartz
From h

30. Rhyod
cone. A

31. Argilli
bedded
Bay

32. Rhyoli
E.) on

33. Granop
N., lon

34. Bytow
53°46.1

35. Andesi
Specim
at an a

36. Hornb
(Fire I

37. Andesi
1926, i

TABLE 1.—ANALYSES OF ROCKS FROM UMNAK AND BOGOSLOF ISLANDS, ALEUTIAN ISLANDS, ALASKA

Abundance of minor elements determined spectrographically.

8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37
											MAJOR OXIDES (Weight per cent)																			
SMB RLH LMS	FMB WWH	WWH	WWH	WWB LMS	FMB WWH	FMB	FMB WWH	FMB WWH	SMB RLH LMS	SMB RLH LMS	FMB	MKC	WWB LMS	WWB LMS	WWB LMS	WWB LMS	WWB LMS	SMB RLH LMS	SMB RLH LMS	SMB RLH LMS	SMB RLH LMS	SMB RLH LMS	WWB LMS	SMB RLH LMS	WWB LMS	SMB RLH LMS	WWB LMS	WWB LMS	TMC	TMC
51.4	52.23	52.43	52.58	53.1	54.35	54.41	54.80	57.40	63.5	65.6	72.36	72.77	54.3	56.0	57.4	57.5	60.6	61.3	61.4	64.4	65.6	65.6	68.2	75.4	76.0	46.0	61.0	51.54	59.81
14.4	17.04	15.27	14.97	18.4	16.38	15.77	14.90	15.49	16.1	14.1	13.75	13.29	18.0	15.6	18.2	17.4	15.4	16.3	16.5	16.0	14.1	15.6	11.4	13.0	12.8	18.5	18.6	20.31	18.03
6.3	3.22	5.96	3.86	4.5	3.01	4.27	2.88	1.97	1.4	2.0	0.59	1.48	1.7	2.1	2.0	2.2	2.7	2.1	2.6	2.1	4.5	2.1	2.3	0.5	0.0	6.1	3.9	4.64	3.94
1.3	6.60	6.99	8.81	3.2	8.00	6.55	9.00	7.86	5.0	3.8	2.33	1.55	7.8	4.8	5.0	5.2	5.6	6.0	5.4	4.1	2.2	3.4	6.0	0.8	1.3	5.0	1.2	3.56	0.87
1.5	5.54	4.13	4.00	3.4	3.34	4.02	3.00	2.94	1.2	0.49	0.09	0.08	3.8	8.8	4.0	3.8	2.0	1.6	2.2	1.2	2.0	0.71	3.8	0.22	0.09	5.4	1.1	3.16	1.37
8.4	10.39	8.70	8.64	9.3	8.08	8.29	7.15	6.74	3.8	2.8	1.53	1.27	7.9	7.7	7.7	6.9	3.5	5.6	5.6	4.2	2.8	3.4	0.94	1.2	0.91	11.9	6.5	9.55	6.51
4.4	2.97	3.50	3.47	2.8	3.66[b]	3.50	3.83	3.96	4.6	4.3	4.64	4.72	3.4	2.4	2.9	3.6	5.4	4.6	4.3	4.5	6.1	4.2	3.5	3.5	3.5	2.6	4.2	4.29	4.46
1.1	0.78	1.24	1.22	1.4	1.32[b]	1.26	1.23	1.53	2.6	3.0	4.16	4.06	1.0	1.6	1.4	1.7	2.2	1.7	1.6	2.6	1.5	2.8	0.51	3.9	4.7	1.6	2.9	2.47	3.12
..	0.12[b]	0.20	0.20	1.14	0.08[b]	0.09	0.18[b]	0.20[b]	..	..	0.27	0.10	..	..	..	..	1.07	..	..	..	..	0.81	..	..	..	..	..	..	0.08
6.3	0.06[b]	0.11	0.07	0.54	0.04[b]	0.02	0.09[b]	0.09[b]	−0.08[d]	3.4	0.03	0.02	−0.20[d]	0.03	0.20	0.24	0.31	−0.51[e]	−0.29[e]	−0.11[d]	1.4	0.31	3.1	2.2	0.48	−0.29[d]	0.08	0.34	0.16
..	..	..	..	0.65	..	..	..	..	..	..	..	..	..	..	..	..	0.07	..	..	..	..	0.07	..	..	..	..	..	..	..
0.93	1.24	1.48	1.80	1.0	1.25	1.56	2.48	1.54	0.81	0.64	0.21	0.21	1.2	0.66	0.90	1.2	1.0	1.0	1.1	0.82	0.75	0.78	0.72	0.16	0.18	1.4	0.53	0.32	0.53
2.4	0.25[b]	0.36	0.38	0.30	0.33[b]	0.25	0.65	0.32[b]	0.23	0.16	0.06	0.06	0.32	0.17	0.16	0.4	0.43	0.42	0.18	0.23	0.20	0.27	0.08	0.06	0.02	0.36	0.28	0.57	0.38
0.23	0.19[b]	0.17	0.21	0.14	0.15[b]	0.21	0.25	0.18[b]	0.22	0.18	0.09	0.10	0.18	0.12	0.10	0.10	0.15	0.20	0.17	0.16	0.11	0.12	0.24	0.02	0.02	0.17	0.16	0.32	0.18
98.66	100.63	100.54	100.21	99.85	99.99	100.20	100.44	100.22	99.46	100.47	100.11	99.71	99.60	99.98	99.96	100.24	100.43	100.82	101.05	100.31	101.26	100.17	100.79	100.96	100.0	99.63	100.45	101.07	99.44
2.55	2.94	2.89	2.93	2.78	2.88	2.8[e]	2.92	2.83	2.58	2.45	2.4[e]	2.4[e]	2.79	2.88	2.81	2.84	2.76	2.59	2.67	2.59	2.71	2.48	2.71	2.46	2.61	2.86	2.60	..	..
											MINOR ELEMENTS[f]																		
		KJM	KJM	ATM			KJM	KJM			KJM		ATM		ATM	AC	ATM					ATM		ATM	AC	ATM	AC		
		.007	.008	.004			.01	.01			.02		.007		.005	.007	.008					.02		.008	.02	.006	.009		
		.0007	.001	.001			.001	.0007			.001		.001		.0008	.002	.001					.001		.0008	.0009	.001	.002		
		..	..	..			..	..			..		..		..	..	..					..		..	..	..	..		
		..	..	..			..	..			..		..		..	.002	..					..		..	.0003	..	.0002		
		.005	.006	.002			.007	.007			.01		.004		.002	.004	.004					.006		.002	.003	.004	.0009		
		..	..	..			..	..			..		..		..	.0003	..					..		..	.0003	..	.00009		
		..	..	..			..	..			.006		.002		.002	.002	..					.002		.008	..	..	.001		
		..	..	..			..	..			.0005		..		..	..	..					..		..	..	..	..		
		.1	.1	.05			.06	.1			.02		.05		.03	.06	.03					.03		.02	.01	.08	.07		
		.03	.03	.05			.03	.05			.09		.02		.05	.02	.06					.06		.07	.05	.06	.1		
		..	..	..			.001	..			.005		..		..	..	..					..		..	..	..	..		
		..	..	..			..	..			..		..		..	.0006	..					..		..	..	..	.0002		
		.03	.04	.007			.01	.008			.004		.009		.006	.005	.003					.003		.0008	.0008	.01	.0007		
		.03	.05	.02			.02	.02			.001		.02		.01	.02	.02					.002		..	.002	.03	.008		
		.004	.01	.007			..	.003			.0003		.0007		.007	.01	.0003					..		..	.0001	.002	.0002		
		.002	.004	.003			..	.001			..		.001		.004	.003	.001					..		..	.0001	.003	.0001		
		.002	.003	.002			.002	.002			..		.002		.002	.003	.001					.0007		..	.0001	.002	.0008		
		..	..	..			..	..			..		..		..	.002	..					..		..	..	..	..		
		.2107	.252	.146			.141	.2017			.1618		.1167		.1208	.1409	.1283					.1266		.1096	.0876	.1980	.1932		
											C. I. P. W. NORMS (Weight Per Cent)																			
4.6	2.8	4.9	3.7	10.1	4.8	6.7	7.7	8.3	14.5	21.5	25.2	27.0	5.1	6.0	11.3	9.2	9.5	12.6	14.0	16.5	17.2	21.3	35.2	37.5	35.0	..	12.4	..	11.1
..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	0.2	3.8	1.0	0.5	..	..	..	..
6.7	4.5	7.3	7.2	8.3	7.8	7.5	7.2	8.9	15.6	17.8	24.6	24.5	6.1	9.4	8.3	10.0	12.8	10.0	9.5	15.6	8.9	16.7	2.8	22.8	27.8	9.4	17.2	14.5	18.3
37.2 / 16.1	53.3 / 25.1 / 30.8	55.9 / 29.6 / 22.3	51.9 / 29.4 / 21.7	51.1 / 23.6 / 33.4	57.0 / 30.9 / 24.7	55.6 / 29.6 / 23.6	53.2 / 32.5 / 19.7	52.2 / 33.5 / 20.0	53.5 / 38.8 / 15.6	54.4 / 36.1 / 10.3	46.4 / 39.3 / 4.4	43.7 / 39.8 / 2.8	28.8 / 30.6	59.4 / 20.4 / 27.0	47.4 / 24.6 / 32.3	56.9 / 30.4 / 26.4	56.8 / 45.6 / 11.4	57.0 / 38.8 / 18.9	57.7 / 36.2 / 21.1	57.3 / 38.3 / 15.6	53.9 / 51.4 / 6.7	58.1 / 35.6 / 14.7	50.3 / 29.3 / 3.9	33.2 / 29.3 / 5.6	34.9 / 29.3 / 4.2	33.5	18.7 / 33.9	52.6 / 35.6 / 23.1	58.7	29.6 / 28.9	58.5 / 34.6 / 21.4
..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	1.9	..	3.6	..
4.2	7.9	7.7	7.8	3.0	5.5	6.6	4.7	4.8	1.0	1.2	1.2	1.2	2.9	4.4	2.2	2.1	1.2	2.6	2.4	1.6	2.7	..	..	..	..	9.6	3.0		
3.6 / 7.8	4.8 / 15.4	4.7 / 15.0	3.7 / 15.4	2.4 / 5.7	2.4 / 11.0	3.9 / 12.9	1.9 / 9.4	1.9 / 9.5	0.3 / 2.1	0.3 / 2.4	0.1 / 2.5	0.17 / 2.47	1.3 / 5.8	3.1 / 8.4	1.3 / 4.3	1.2 / 4.1	0.5 / 2.4	0.9 / 5.2	1.2 / 4.8	0.7 / 3.2	2.3 / 5.0	..	..	..	..	7.4 / 18.2	2.6 / 5.6	12.0	6.5
..	2.7	2.6	3.9	0.3	3.1	2.4	2.8	2.8	0.8	0.9	1.2	1.1	1.6	0.9	0.8	0.8	0.7	1.7	1.2	0.9	..	..	..	..	..	1.2	..		
0.1 / 0.1	9.0 / 14.1	5.6 / 8.8	6.2 / 12.8	6.1 / 6.8	5.9 / 13.3	6.1 / 10.0	5.5 / 13.3	5.4 / 13.2	2.7 / 8.9	0.9 / 4.5	0.2 / 2.6	0.03 / 0.25	8.2 / 9.5	17.7 / 18.9	24.2 / 8.7	14.0 / 8.3	13.3 / 4.5	10.4 / 3.1	9.3 / 4.3	9.3 / 2.3	6.1 / 2.7	2.7	1.8 / 3.4	5.2 / 9.5	17.8 / 0.5	1.3 / 0.2	2.2	..	0.1	0.1	..	0.4
..	5.1	3.2	6.6	0.4	7.4	3.9	7.8	7.8	6.2	3.6	2.4	0.22	9.5	5.3	5.3	5.0	..	5.9	6.2	5.0	..	..	8.3	0.8	2.0	..	..	..	..
..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	4.3	5.0	3.5	..
..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	0.7	..		
2.3	4.7	8.7	5.6	6.5	4.4	6.2	4.2	2.9	2.1	3.0	0.9	2.1	2.6	3.0	3.0	3.3	3.9	3.0	3.7	3.0	5.3	3.0	3.3	0.7	..	8.8	2.5	6.7	2.3
1.7	2.4	2.8	3.4	2.0	2.4	2.9	4.7	2.9	1.5	1.2	0.4	0.4	2.4	1.2	1.5	2.3	2.0	2.0	2.1	1.5	1.4	1.5	1.4	0.3	0.5	2.7	1.1	0.6	0.9
10.1	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	0.8	..	..	..	..	..	1.9	..	2.2
5.7	0.6	0.8	0.9	0.7	0.8	0.6	1.6	0.8	0.3	0.3	0.1	0.1	0.7	0.3	0.3	1.0	1.0	1.0	0.3	0.5	0.4	0.7	0.3	0.2	0.07	0.8	0.7	1.3	1.0
..	..	..	..	1.6	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	..	0.2	..	..	..	..	..	..	..
92.3	100.4	100.2	100.1	98.4	100.1	100.0	100.3	100.0	99.4	97.1	100.0	99.42	99.8	99.9	99.6	100.0	99.0	100.8	101.0	100.3	99.8	99.1	97.8	98.7	99.57	99.4	100.2	100.7	98.7
											NIGGLI VALUES																			
75	129	134	136	148	148	146	155	170	236	286	372	384	148	141	167	169	207	203	200	240	248	270	286	471	473	105	204		
29	25	23	23	30	26	25	25	27	35	36	42	42	29	23	31	30	31	32	31½	35	31½	38	28	48	47	24½	37		
24	40½	42	42	32	38	40	41	37½	27	24½	13	13	37	47½	34	35	33½	30	32	26	31½	23	52½	8	7½	38	19		
31	27	24	24	28	24	24	22	21½	15	13	8	7	23	21	24	22	13	20	19½	17	11	15	4	8	6	29	24		
17	7½	11	11	10	12	11	12	14	23	26½	37	38	11	8½	11	13	22½	18	17	22	26	24	15½	36	39½	8½	20		
0.14	0.15	0.19	0.19	0.25	0.19	0.19	0.17	0.20	0.27	0.32	0.37	0.37	0.17	0.30	0.24	0.24	0.21	0.20	0.20	0.28	0.14	0.31	0.08	0.42	0.47	0.29	0.31		
0.32	0.50	0.37	0.36	0.45	0.35	0.40	0.31	0.35	0.25	0.13	0.05	0.05	0.42	0.70	0.51	0.48	0.30	0.26	0.34	0.26	0.36	0.18	0.45	0.23	0.10	0.48	0.25		
7	−1	−10	−8	8	0	2	7	14	44	80	124	134	4	7	23	17	17	27	32	50	44	74	124	227	215	−29	24		

lagioclase-olivine basalt flow (46 ABy 92) of Mt. Tulik (lat. 53°22.4′ N., long. 57°04.3′ E.)

alagonitized tuff cone (46 ABy 132) of early postcaldera pyroclastic rocks within kmok Caldera (lat. 53°26.4′ N., long. 168°05.0′ E.), 1 mile northeast of cone D Pl. 1). Specimen collected from palagonitized bed near summit.

northite-augite-olivine basalt flow (46 AFr 28) of Ashishik basalt (lat. 53°27.7′ N., ng. 167°55.5′ E.), 3 miles N. 25° W. of Umnak Airfield. Specimen collected from wer middle part of flow.

northite basalt flow (46 ABy 609) near Inanudak Bay (lat. 53°17.7′ N., long. 8°21.3′ E.), from sea cliff S. 25° E. of cinder cone

northite basalt flow (46 AHp 68) extruded in December 1945 from cone A (Pl. 1) t. 53°24.7′ N., long. 168°11.0′ E., 6000 feet northwest of cone)

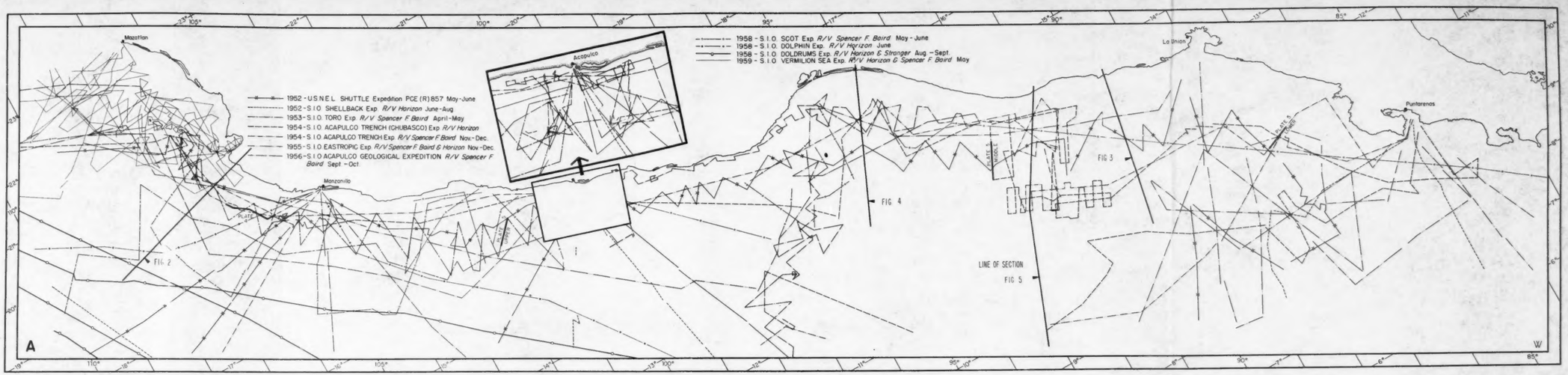
ytownite-augite-olivine basalt flow (46 ABy 29), a subaerial flow of cone D (Pl. 1) t. 53°25.6′ N., long. 168°05.8′ E., south-southeast slope, altitude 1900 feet)

lagonitized scoria bed (45–73) from Ashishik basalt (lat. 53°21.2′ N., long. 168° .0′ E.)

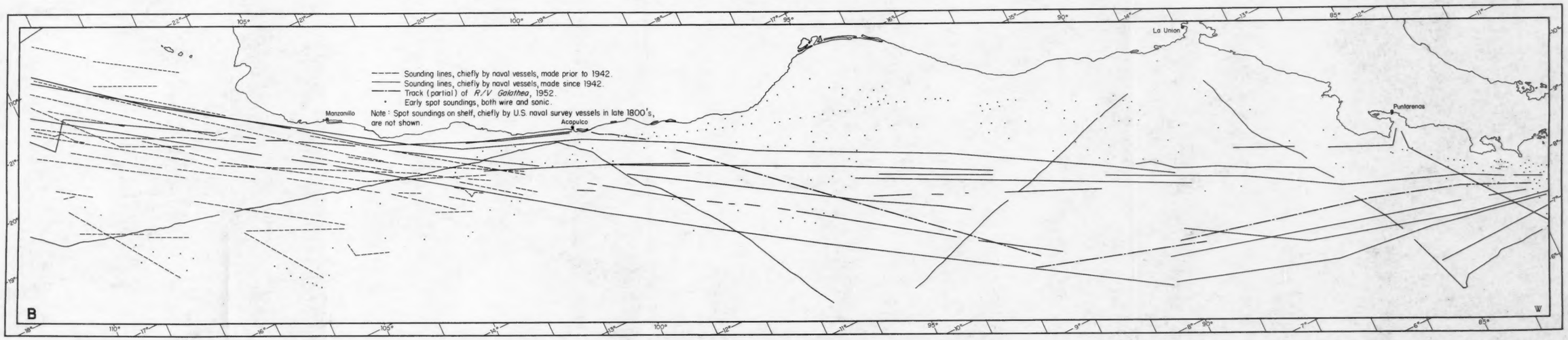
gite basalt flow (46 ABy 70) of Ashishik basalt in lower part of Okmok Caldera ll (lat. 53°27.3′ N., long. 168°05.5′ E.)

ow no. 2 (46 ABy 45) of Crater Creek basalt (lat. 53°28.5′ N., long. 168°05.0′ E.) northwest wall of Crater Creek gorge

11. Flow no. 11 (46 ABy 54) of Crater Creek basalt (lat. 53°28.7′ N., long. 168°04.6′ E.) in northwest wall of Crater Creek gorge
12. Hydrothermally altered plagioclase basalt (47 ABy 44) from volcanic rocks of central Umnak. From cliff just south of Air Force hut near Inanudak Bay (lat. 53°17.1′ N., long. 168°19.8′ E.)
13. Lower welded andesitic agglomerate bed (46 ABy 13) in Okmok volcanics exposed in upper part of north wall (lat. 53°28.3′ N., long. 168°08′ E.) of Okmok caldera. Sample taken to exclude most foreign lithic fragments.
14. Older basaltic andesite flow (46 ABy 6) from cone B (Pl. 1) in Okmok Caldera. Near front of flow (lat. 53°27.6′ N., long. 168°07′ E.)
15. Aphyric andesite pipe (46 ABy 63) of vitreous andesite. Near top of pipe (lat. 53°28.0′ N., long. 168°05.4′ E.) at intersection of southeast wall of Crater Creek gorge and Okmok Caldera
16. Upper welded andesitic agglomerate bed (46 ABy 15) in Okmok volcanics exposed in upper part of north wall (lat. 53°28.3′ N., long. 168°08′ E.) of Okmok Caldera. Sample taken to exclude most xenolithic fragments.
17. Vitreous latite fragment (46 AFr 2) in Okmok volcanics. From bank of Crater Creek (lat. 53°32.6′ N., long. 167°59.2′ E.) near Cape Tanak
18. Rhyodacite pumice lapilli bed (47 ABy 26) near base of Okmok volcanics. Sea cliff (lat. 53°33.5′ N., long. 168°04.9′ E.) 0.6 mile S. 10° E. of Ashishik Point
19. Rhyolite obsidian band (46 ABy 30a) in rhyolite from northeastern Umnak (lat. 53°29.5′ N., long. 168°10.3′ E.)
20. Felsitic rhyolite facies (48 Um 11) in east part of rhyolite from northeastern Umnak (lat. 53°29.4′ N., long. 168°09′ E.)
21. Basaltic andesite flow (47 ABy 73) of Mount Vsevidof from Twinlava Point (lat. 53°11.1′ N., long. 168°47.7′ E.)
22. Quartz-bearing olivine andesite flow (47 ABy 10). North-facing cliff (lat. 53°13.6′ N., long. 168°21.9′ E.), south of Inanudak Bay
23. Hypersthene-bearing labradorite andesite flow (47 ABy 9) of Mount Recheschnoi (lat. 53°13.8′ N., long. 168°22.0′ E.), south of Inanudak Bay
24. Quartz diorite (47 ABy 41) facies of plutonic rocks (lat. 53°16.1′ N., long. 168°34.9′ E.), 1¼ miles S. 70° E. of Cape Ilmalianuk
25. Albitized diorite (47 ABy 34) in albitized sedimentary and igneous complex. From south shore line (lat. 52°51.8′ N., long. 169°00.1′ E.) 1½ miles east of Idaliuk Point
26. Andesitic glass clot (47 ABy 72) in dark scoria bed of Mount Vsevidof cone. Specimen collected at an altitude of 4000 feet on north slope of cone (lat. 53°07.7′ N., long. 168°41.3′ E.)
27. Hypersthene-bearing andesite flow (47 ABy 57) of Mount Recheschnoi, south slope (lat. 53°05.3′ N., long. 168°28.5′ E.)
28. Historic(?) latite flow (47 ABy 36) of Mount Vsevidof. Front of flow (lat. 53°07.7′ N., long. 168°46.6′ E.) on west slope of Mount Vsevidof
29. Quartz keratophyre (47 ABy 29) in albitized sedimentary and igneous complex. From headland (lat. 53°02.4′ N., long. 168°30.0′ E.) east of Amos Bay
30. Rhyodacite pumice bed (47 ABy 67) that forms uppermost bed of Mount Vsevidof cone. Altitude 1700 feet on north slope of cone (lat. 53°09.7′ N., long. 168°41.7′ E.)
31. Argillite bed (47 ABy 54) in albitized sedimentary and igneous complex. From bedded sequence (lat. 53°07.7′ N., long. 168°22.4′ E.), 1 mile southwest of Russian Bay
32. Rhyolite dome (47 ABy 28). From middle dome (lat. 53°10.2′ N., long. 168°25.1′ E.) on west side of valley that drains into Russian Bay
33. Granophyre dike (47 ABy 39A) that cuts plutonic rocks. From sea cliff (lat. 53°16.2′ N., long 168°36.9′ E.) three-quarters of a mile southwest of Cape Ilmalianuk
34. Bytownite-salite-hornblende basalt dome (47 ABy 100) extruded in 1927 (lat. 53°46.1′ N., long. 168°02.5′ E.)
35. Andesine-hornblende andesite dome (47 ABy 103) (Old Bogoslof) extruded in 1796. Specimen collected in saddle of Castle Rock (lat. 53°55.8′ N., long. 168°02.3′ E.) at an altitude of 100 feet
36. Hornblende basalt (Merrill, 1885, p. 32) extruded in 1883 from "New Bogoslof" (Fire Island) (lat. 53°56.5′ N., long. 168°02.5′ E.)
37. Andesine-hornblende andesite dome (Harriman Alaska Exped. No. 146) (Fenner, 1926, p. 706) extruded in 1796. Exact location unknown.

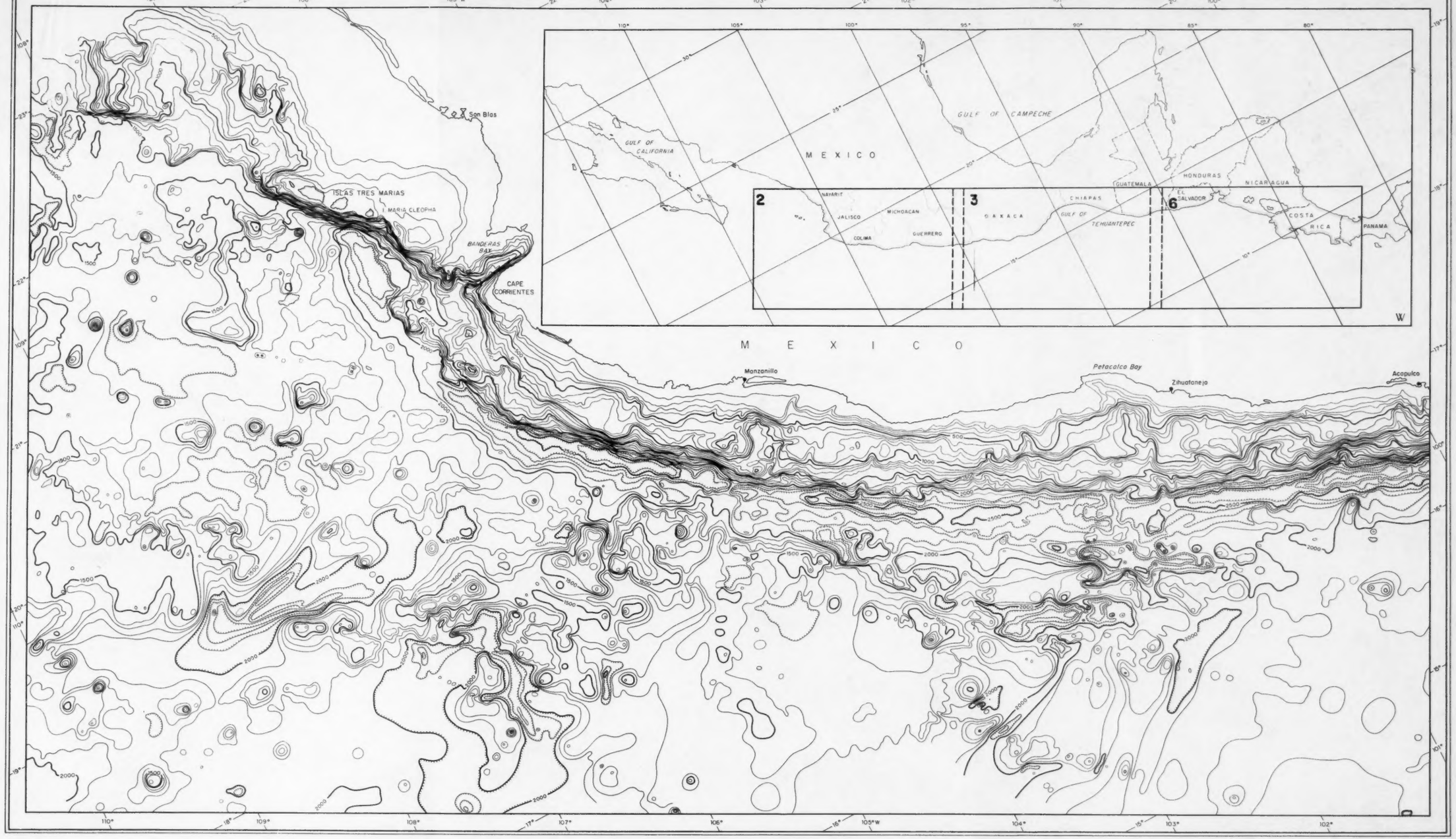

Mazatlan
Acapulco
Manzanillo
La Union
Puntarenas
1952 - U.S.N.E.L. SHUTTLE Expedition PCE(R)857 May-June
1952 - S.I.O. SHELLBACK Exp. R/V Horizon June-Aug.
1953 - S.I.O. TORO Exp. R/V Spencer F.Baird April-May
1954 - S.I.O. ACAPULCO TRENCH (CHUBASCO) Exp. R/V Horizon
1954 - S.I.O. ACAPULCO TRENCH Exp. R/V Spencer F.Baird Nov.-Dec.
1955 - S.I.O. EASTROPIC Exp. R/V Spencer F.Baird & Horizon Nov.-Dec.
1956 - S.I.O. ACAPULCO GEOLOGICAL EXPEDITION R/V Spencer F. Baird Sept.-Oct.
1958 - S.I.O. SCOT Exp. R/V Spencer F. Baird May - June
1958 - S.I.O. DOLPHIN Exp. R/V Horizon June
1958 - S.I.O. DOLDRUMS Exp. R/V Horizon & Stranger Aug. - Sept.
1959 - S.I.O. VERMILION SEA Exp. R/V Horizon & Spencer F. Baird May
FIG 2
FIG 3
FIG 4
LINE OF SECTION
FIG 5
A
W

A. Index map: S.I.O. and N.E. L. exploration of the Middle America Trench


Sounding lines, chiefly by naval vessels, made prior to 1942.
Sounding lines, chiefly by naval vessels, made since 1942.
Track (partial) of R/V Galathea, 1952.
Early spot soundings, both wire and sonic.
Note : Spot soundings on shelf, chiefly by U.S. naval survey vessels in late 1800's, are not shown.
Manzanillo
Acapulco
La Union
Puntarenos
B
W

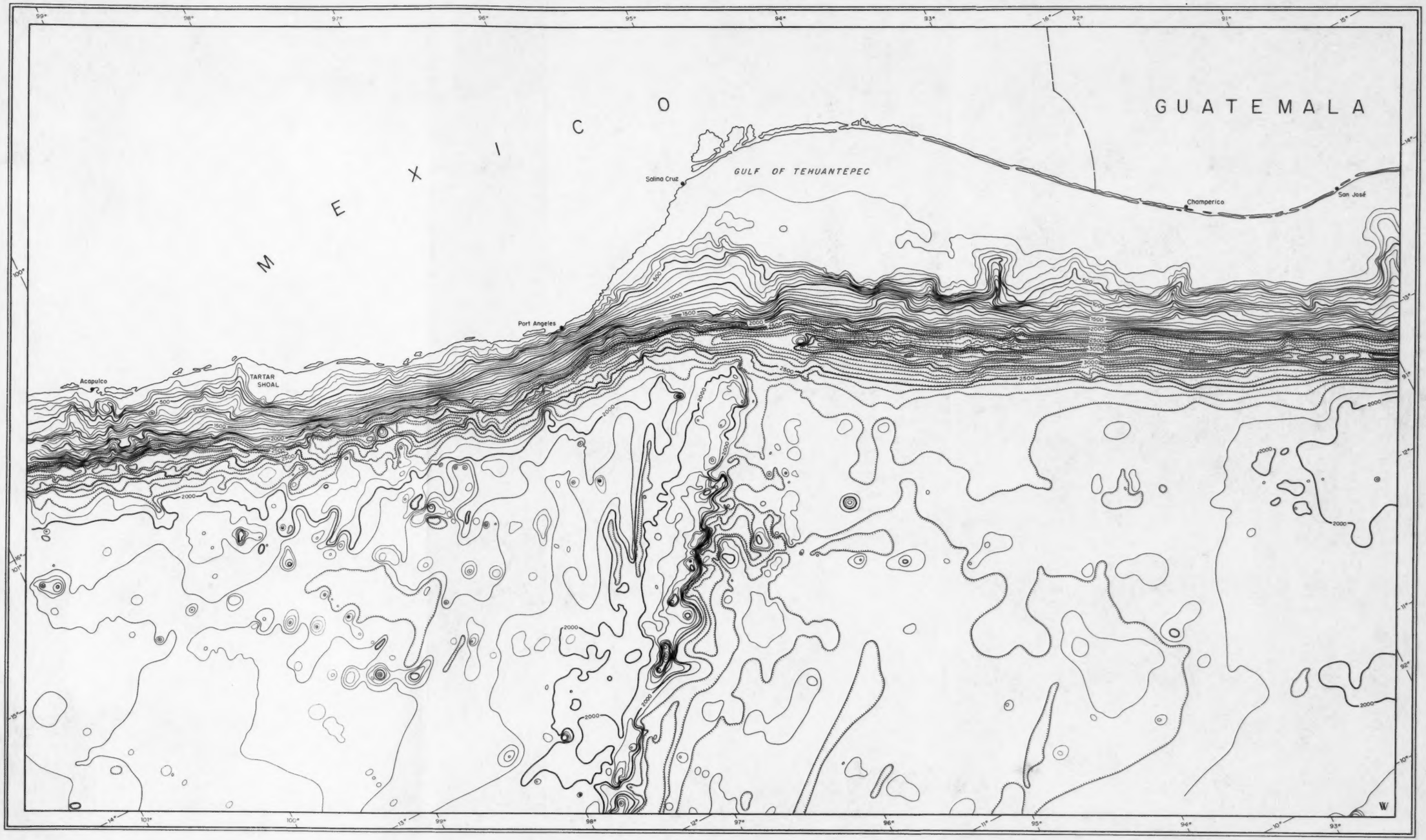
B. Index map: Sounding lines furnished by the U.S. Hydrographic Office

ECHO-SOUNDING LINES USED IN THE PREPARATION OF PLATES 2, 3, AND 6



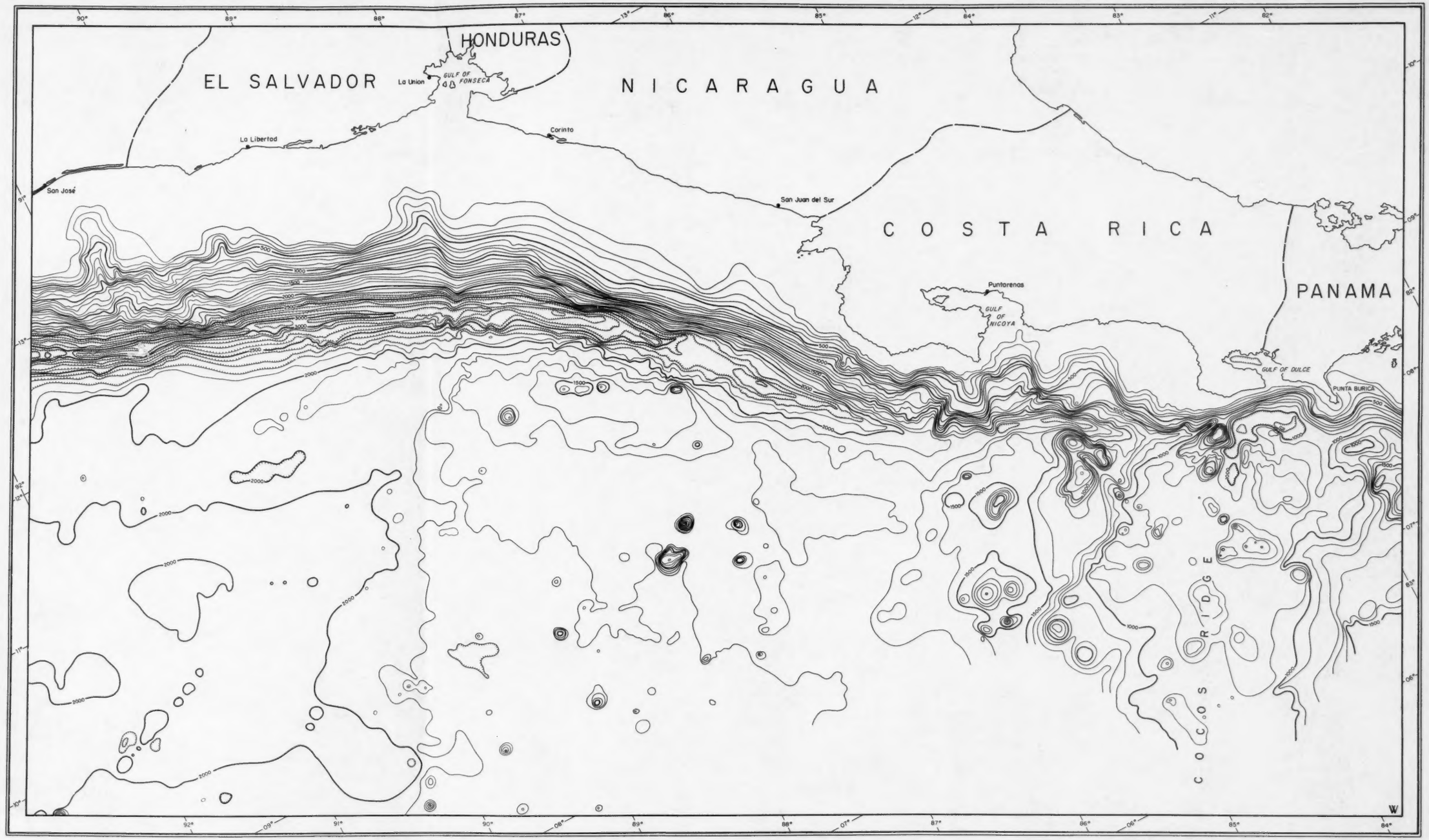
SUBMARINE TOPOGRAPHY, FROM ISLAS TRES MARÍAS TO ACAPULCO, MEXICO

FISHER, PLATE 2
Geological Society of America, volume 72



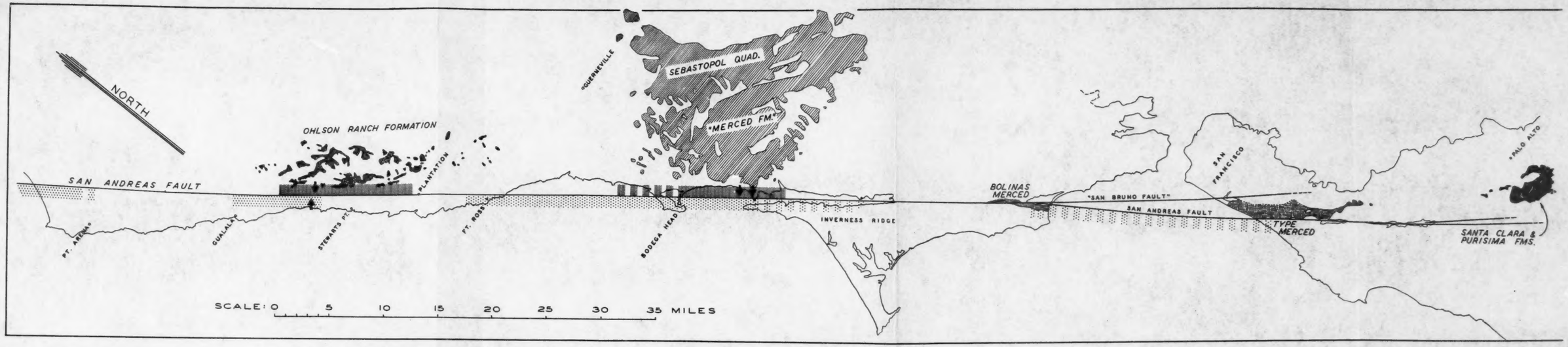
SUBMARINE TOPOGRAPHY, FROM ACAPULCO, MEXICO, TO SAN JOSÉ, GUATEMALA

FISHER, PLATE 3
Geological Society of America, volume 72

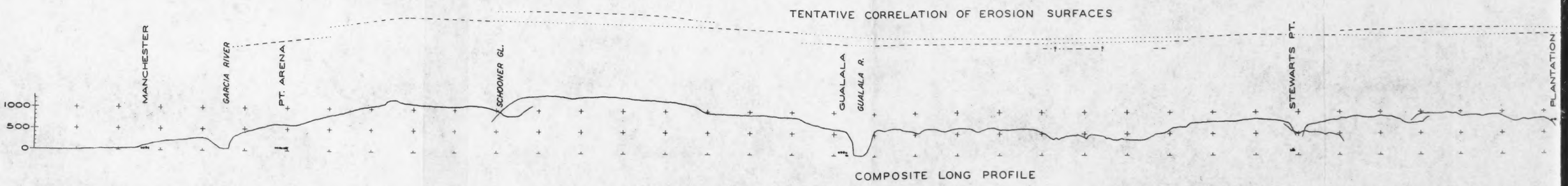


SUBMARINE TOPOGRAPHY, FROM SAN JOSÉ, GUATEMALA, TO PUNTA BURICA, PANAMA

FISHER, PLATE 6
Geological Society of America, volume 72



A. MIDDLE AND UPPER PLIOCENE SEDIMENTS ALONG THE SAN ANDREAS FAULT TRACE BETWEEN PALO ALTO AND POINT ARENA, CALIFORNIA

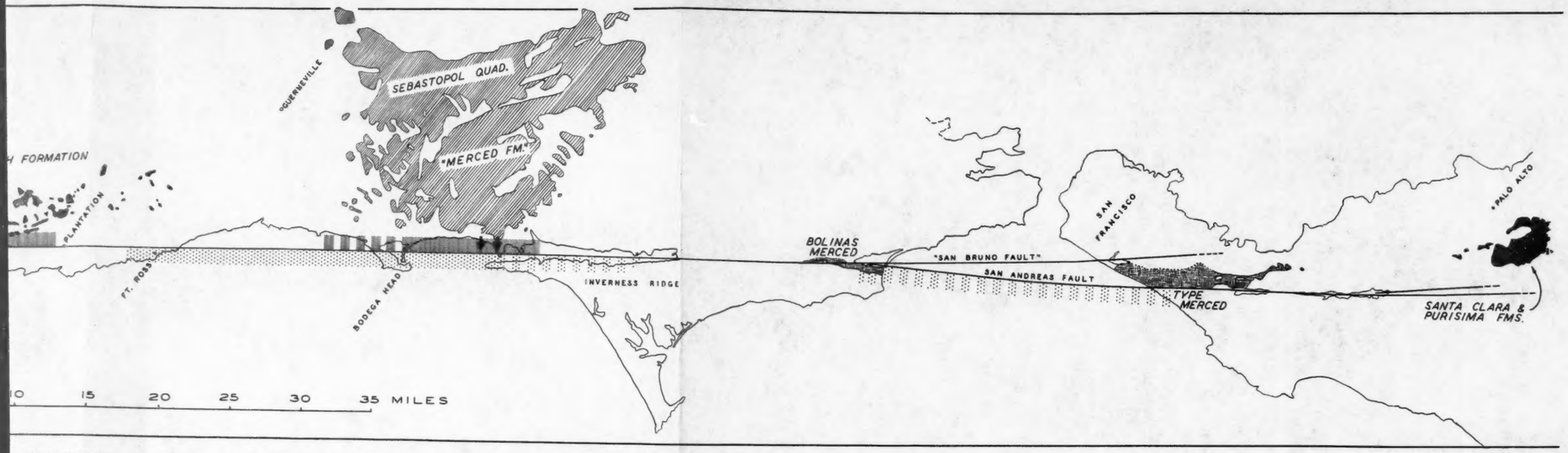


B. COMPOSITE LONG PROFILE AND TENTATIVE CORRELATION OF EROSION SURFACES ON THE SUMMIT OF THE COASTAL RIDGE WEST OF THE SAN ANDREAS FAULT TRACE BETWEEN FORT ROSS AND POIN

Crosses are spaced horizontally at intervals of 1 mile and vertically at intervals of 500 feet. Vertical exaggeration about 5:1. Topographic data from U. S. Geological Survey maps and Army Corps of Engineers maps.

PLIOCENE SEDIMENTS AND EROSION SURFACES NEAR THE SAN ANDREAS FAULT NORTH OF PALO ALTO, CALIF

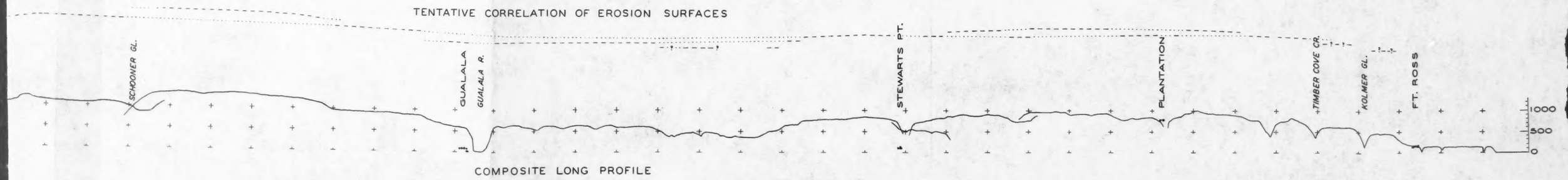
HIGGINS, PLATE 1
Geological Society of America Bulletin, volume 72



AND UPPER PLIOCENE SEDIMENTS ALONG THE SAN ANDREAS FAULT TRACE BETWEEN PALO ALTO AND POINT ARENA, CALIFORNIA

Line labeled "San Andreas fault" represents the 1906 fault trace; the fault-zone trace, not shown, borders this line and encloses most or all of the Bolinas Merced. Marine Pliocene deposits west of the fault trace south of San Francisco are not shown. Geologic data in Palo Alto area from Branner, Newson, and Arnold (1909); type Merced and Bolinas Merced areas from Lawson (1914), modified by Higgins (this paper); Sebastopol "Merced" chiefly from Johnson (1943, p. 624); Ohlson Ranch area from Higgins (1960).

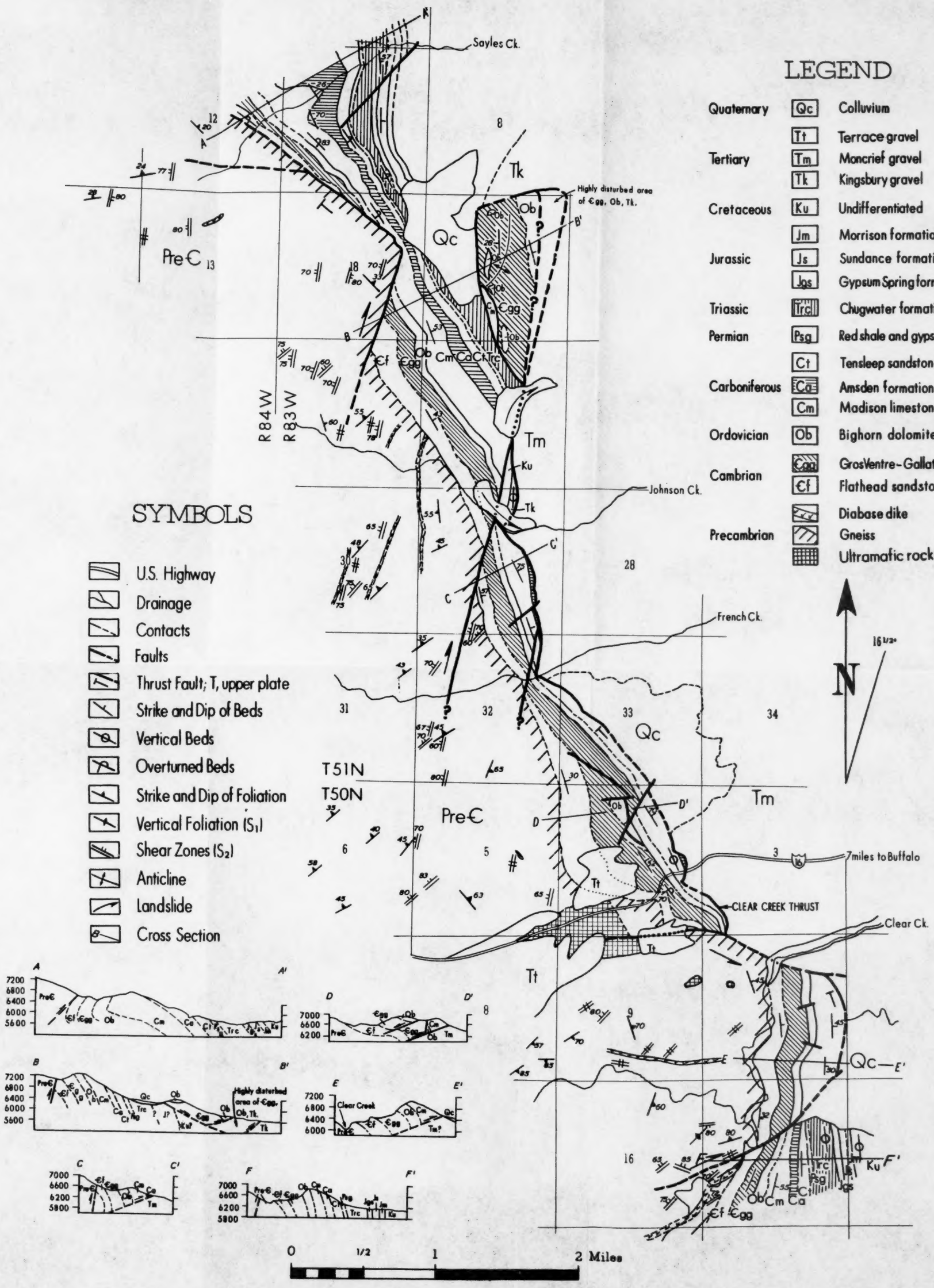
Bands of dots and of wavy lines show areas adjacent to the fault trace north of San Francisco that appear to have been below sea level during middle or late Pliocene time. Dashed bands show areas that may or may not have been submerged during Pliocene time.



PROFILE AND TENTATIVE CORRELATION OF EROSION SURFACES ON THE SUMMIT OF THE COASTAL RIDGE WEST OF THE SAN ANDREAS FAULT TRACE BETWEEN FORT ROSS AND POINT ARENA, CALIFORNIA

Crosses are spaced horizontally at intervals of 1 mile and vertically at intervals of 500 feet. Vertical exaggeration about 5:1. Topographic data from U. S. Geological Survey maps and Army Corps of Engineers maps.

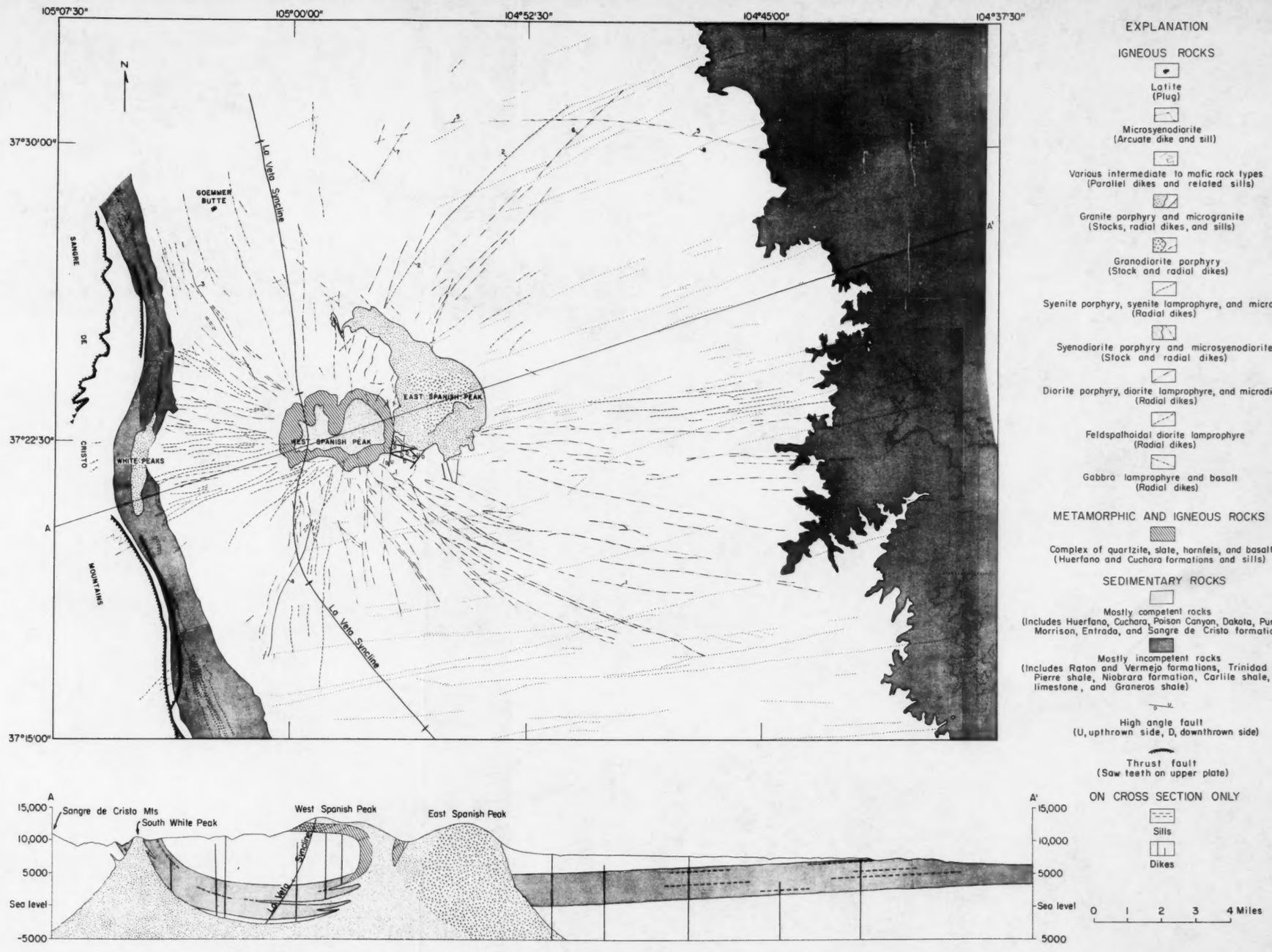
OCENE SEDIMENTS AND EROSION SURFACES NEAR THE SAN ANDREAS FAULT NORTH OF PALO ALTO, CALIFORNIA



GEOLOGIC MAP AND CROSS SECTIONS, CLEAR CREEK THRUST AREA

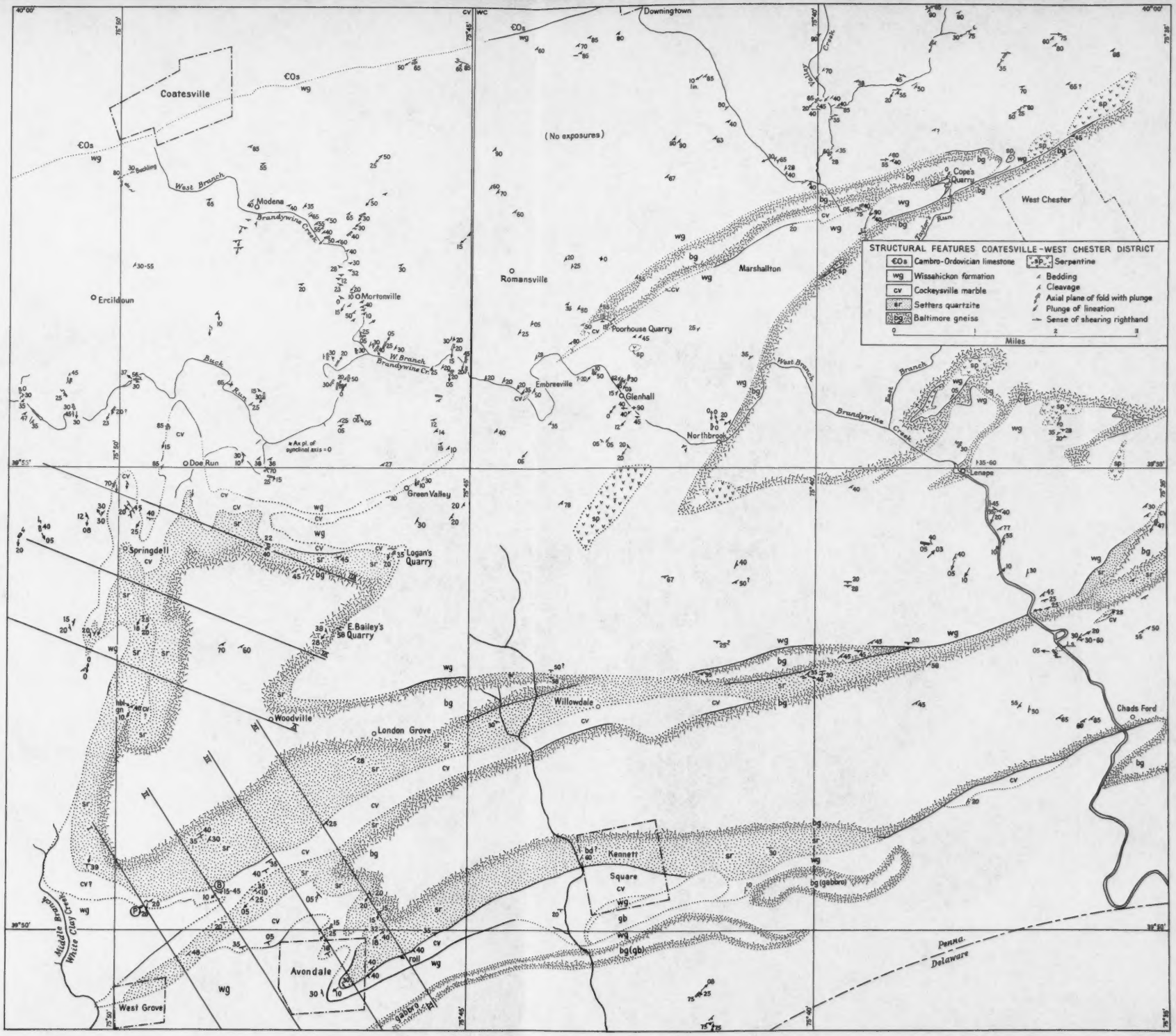
NEAR BUFFALO, WYOMING

HOPPIN, PLATE 1
Geological Society of America Bulletin, volume 72


105°07'30"
105°00'00"
104°52'30"
104°45'00"
104°37'30"
37°30'00"
37°22'30"
37°15'00"
N
GOEMMER BUTTE
La Veta Syncline
SANGRE DE CRISTO MOUNTAINS
WHITE PEAKS
EAST SPANISH PEAK
WEST SPANISH PEAK
A
A'
La Veta Syncline
EXPLANATION
IGNEOUS ROCKS
Latite (Plug)
Microsyenodiorite (Arcuate dike and sill)
Various intermediate to mafic rock types (Parallel dikes and related sills)
Granite porphyry and microgranite (Stocks, radial dikes, and sills)
Granodiorite porphyry (Stock and radial dikes)
Syenite porphyry, syenite lamprophyre, and microsyenite (Radial dikes)
Syenodiorite porphyry and microsyenodiorite (Stock and radial dikes)
Diorite porphyry, diorite lamprophyre, and microdiorite (Radial dikes)
Feldspathoidal diorite lamprophyre (Radial dikes)
Gabbro lamprophyre and basalt (Radial dikes)
METAMORPHIC AND IGNEOUS ROCKS
Complex of quartzite, slate, hornfels, and basalt (Huerfano and Cuchara formations and sills)
SEDIMENTARY ROCKS
Mostly competent rocks (Includes Huerfano, Cuchara, Poison Canyon, Dakota, Purgatoire, Morrison, Entrada, and Sangre de Cristo formations)
Mostly incompetent rocks (Includes Raton and Vermejo formations, Trinidad sandstone, Pierre shale, Niobrara formation, Carlile shale, Greenhorn limestone, and Graneros shale)
High angle fault (U, upthrown side, D, downthrown side)
Thrust fault (Saw teeth on upper plate)
ON CROSS SECTION ONLY
Sills
Dikes
A
A'
15,000
10,000
5000
Sea level
-5000
Sangre de Cristo Mts
South White Peak
West Spanish Peak
East Spanish Peak
La Veta Syncline
0 1 2 3 4 Miles

WEST SPANISH PEAK AND ASSOCIATED DIKE SWARM; DIAGRAMMATIC STRUCTURE CROSS SECTION THROUGH THE WHITE PEAKS, WEST SPANISH PEAK, AND EAST SPANISH PEAK STOCKS

JOHNSON, PLATE 1
Geological Society of America Bulletin, volume 72

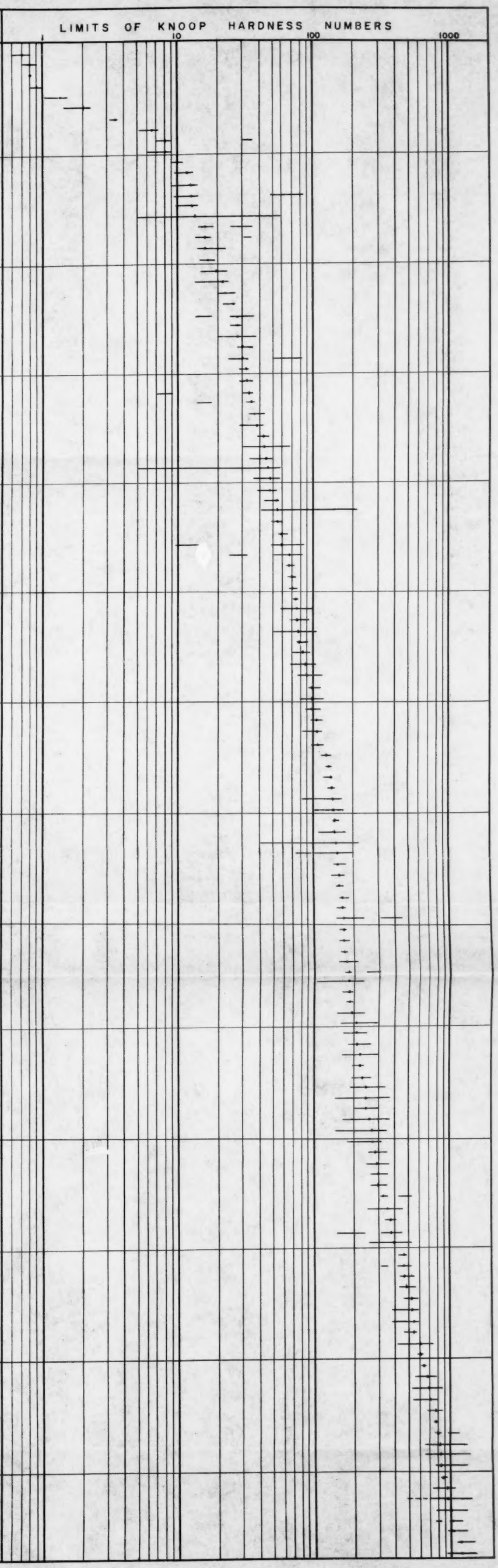


STRUCTURAL FEATURES, COATESVILLE—WEST CHESTER DISTRICT, PENNSYLVANIA

McKINSTRY, PLATE 1
Geological Society of America Bulletin, volume 72

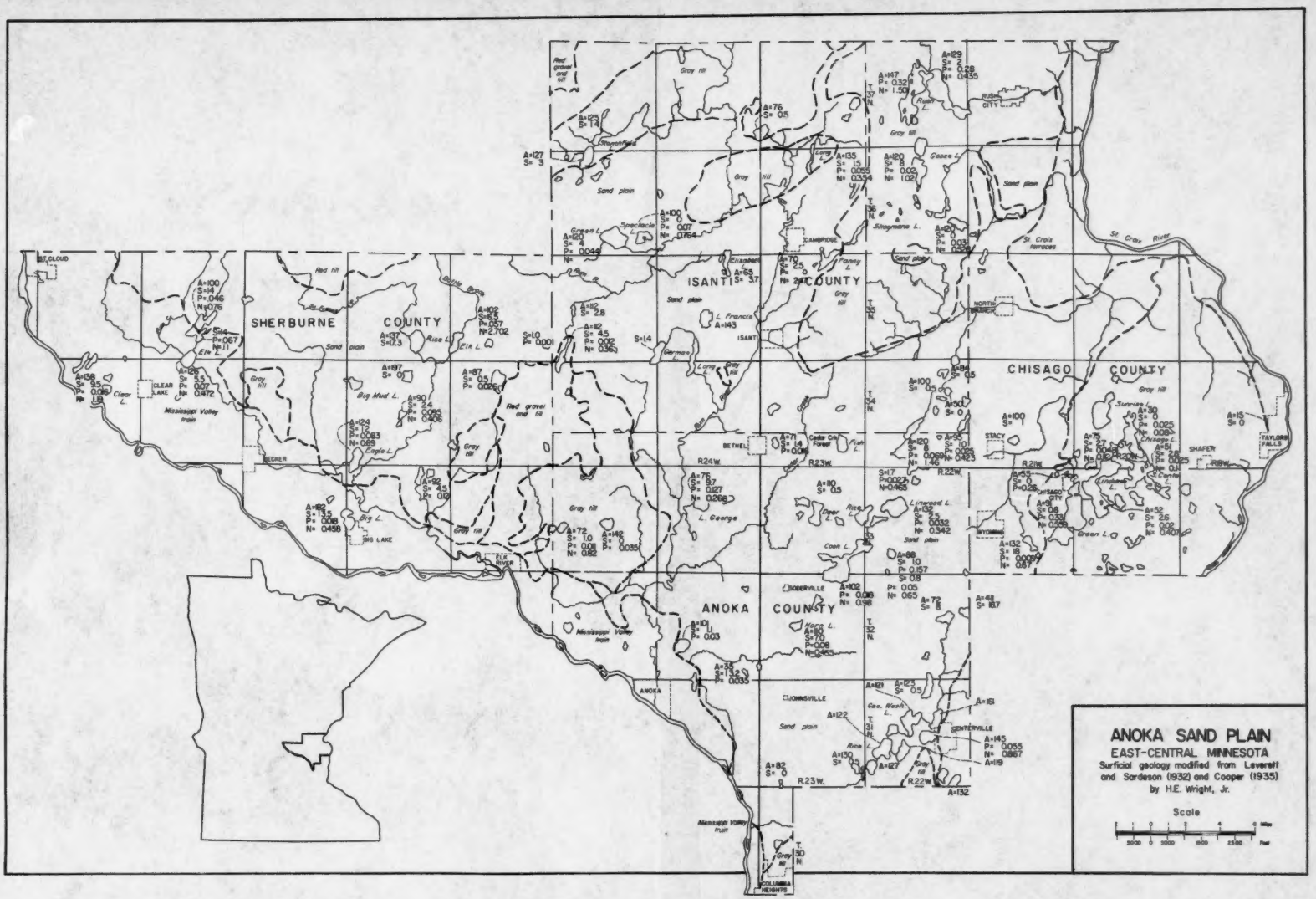
TABLE 5. LIMITS AND AVERAGE KNOOP NUMBERS (LOGARITHMIC SCALE)

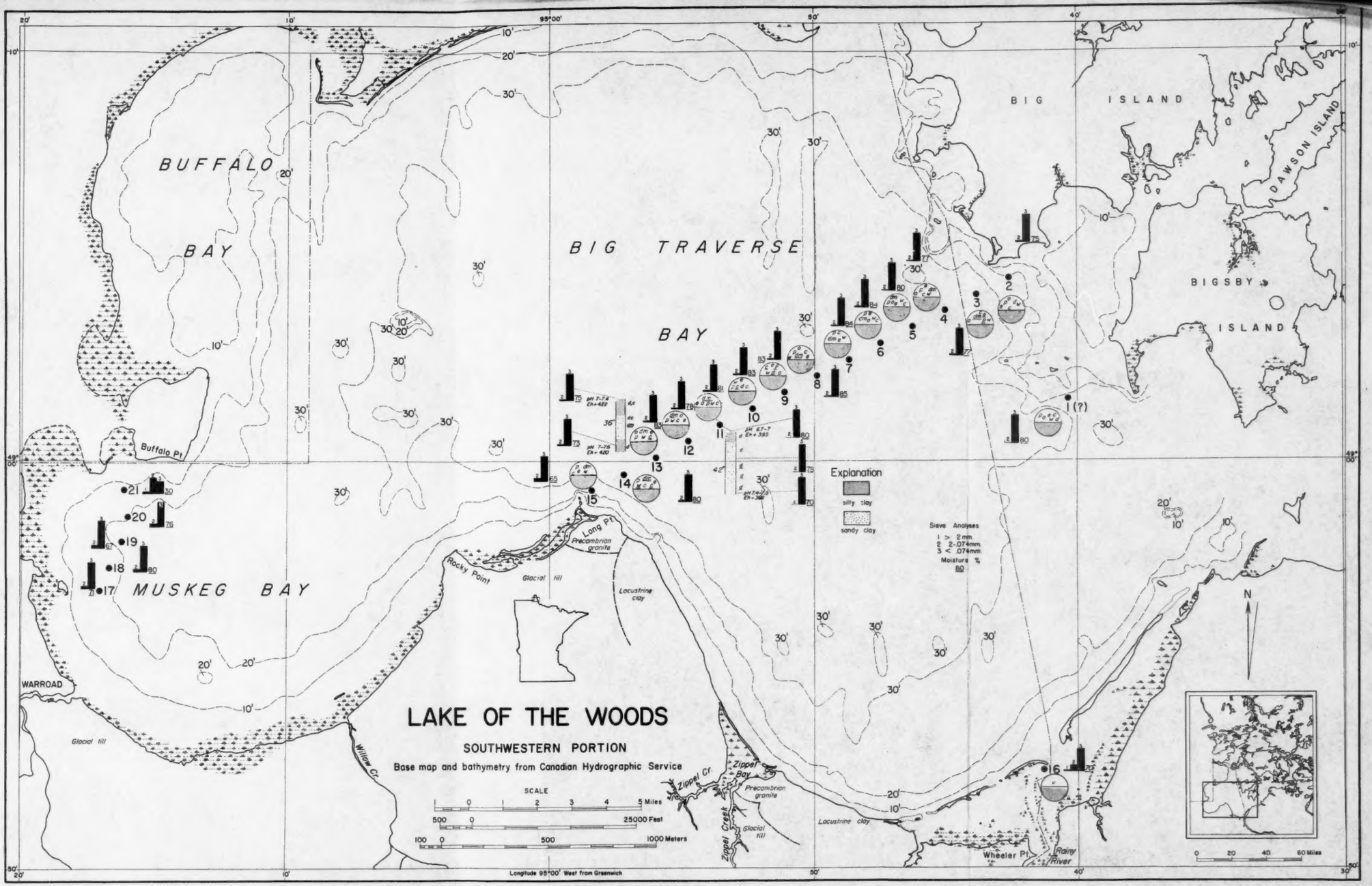
MINERALS	AVERAGE KNOOP NO.
Bismuth	0.7
Kermesite	0.8
Krennerite	0.8
Hessite	0.9
Altaite	1.2
Matildite	2.0
Arsenic	3.3
Tungstenite	6.4
Aramayoite'	8
Argyrodite	8
Silver-arsenian	10
Silver-antimonian	11
Silver-	12
Molybdenite'	12
Stephanite	12
Graphite'	12.5
Chalcophanite'	16
Cylindrite'	16
Melonite	16
Coloradoite	17
Guanajuatite	19
Jamesonite	21
Aguilarite	22
Argentite	26
Chalcophanite'	28
Livingstonite	28
Rickardite	28
Andorite	30
Nagyagite'	30
Clausthalite	31
Klaprothite	32
Aramayoite'	33
Cylindrite'	33
Miargyrite	36
Tiemannite	38
Alabandite	41
Schapbachite	42
Luzonite	43
Graphite'	44
Realgar	47
Baumhauerite	47
Chalcostibnite	51
Teallite'	53
Tellurium	54
Chalcocite	58
Molybdenite'	60
Nagyagite'	66
Willyamite ?	67
Digenite	70
Umangite'	70
Copper	73
Domeykite	75
Galena	75
Cuprodescloizite	76
Gold	79
Benjaminite	82
Geocronite	87
Antimony	89
Aikinite	97
Boulangerite	97
Bornite	98
Covellite	101
Pararammelsbergite	102
Pyrargyrite	103
Pearceite	109
Fuloppite	112
Rezbanyite	116
Wittichenite	117
Guitermanite	125
Sternbergite	127
Klockmannite	131
Platinum refined	132
Teallite'	133
Famatinite	138
Sartorite	139
Rathite	145
Kallilite	146
Bismuthinite	148
Meneghinite	148
Enargite'	154
Semseyite	155
Plagionite	156
Bournonite	158
Skutterudite	167?
Umangite'	167
Stibnite	169
Chalcopyrite	175
Freislebenite	179
Tennantite	190
Sylvanite	202
Emplectite	202
Awaruite	206
Sphalerite	209
Troilite	210
Algodonite	218
Delafossite	226
Goldfieldite	228
Cuprite	229
Millerite	250
Heterolite	264
Schirmerite	266
Pyrrhotite	281
Pentlandite	294
Ullmannite	295
Tetrahedrite	299
Huebnerite'	313
Colusite	331
Wolframite	370
Enargite'	378
Mottramite ?	393
Niccolite	460
Huebnerite'	463
Quartz	467
Goethite	476
Breithauptite	518
Manganite	520
Magnetite	528
Hausmannite	548
Maucherite	600
Carrollite	604
Sperrylite	633
Uraninite	670
Pyrite	685
Hematite	700
Franklinite	770
Glaucodot	770
Polydymite'	800
Arsenopyrite	814
Chloanthite	830
Smaltite	830
Braunite	886
Marcasite	933
Ilmenite	937
Martite	998
Polydymite'	1030
Corynite	1040
Chromite	1100
Berthierite	1140



[1] Knoop-hardness data for one of two crystallographic orientations

ROBERTSON, TABLE 5
Geological Society of America Bulletin, volume 72

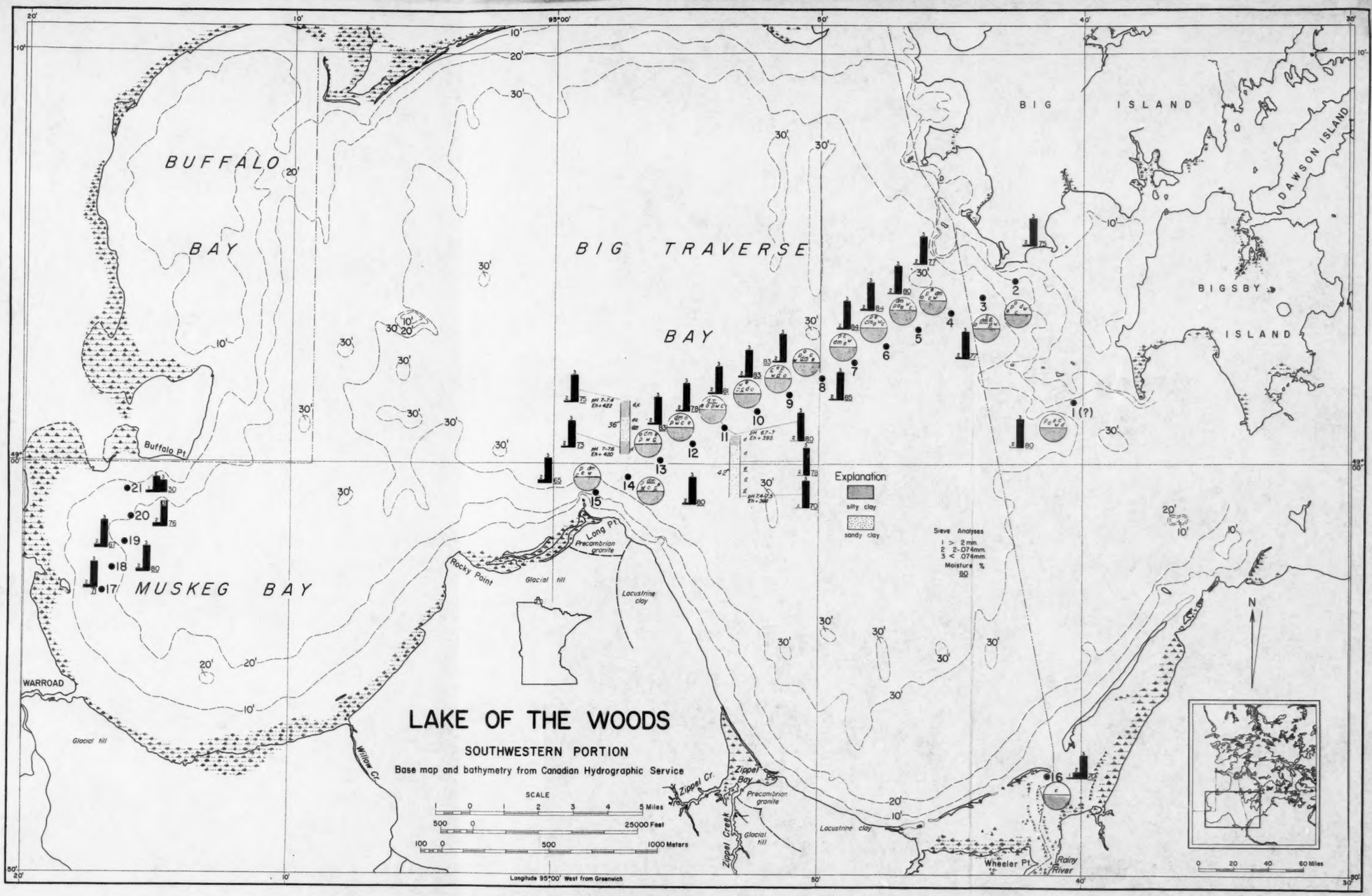

ANOKA SAND PLAIN
EAST-CENTRAL MINNESOTA
Surficial geology modified from Leverett and Sardeson (1932) and Cooper (1935) by H.E. Wright, Jr.
Scale
SHERBURNE COUNTY
ISANTI COUNTY
CHISAGO COUNTY
ANOKA COUNTY
ST. CLOUD
CLEAR LAKE
BECKER
BIG LAKE
ELK RIVER
CAMBRIDGE
ISANTI
NORTH BRANCH
RUSH CITY
BETHEL
STACY
CHISAGO CITY
SHAFER
RUSH
TAYLORS FALLS
ROGERSVILLE
ANOKA
JOHNSVILLE
CENTERVILLE
COLUMBIA HEIGHTS
Red gravel and till
Gray till
Sand plain
Mississippi Valley train
St. Croix terraces
St. Croix River
Rice L.
Elk L.
Big Mud L.
Eagle L.
Fish L.
Cedar Creek Power
L. George
Coon L.
Linwood L.
Chisago L.
Green L.
Typha L.
Sunrise L.
Fanny L.
Spectacle L.
Silogreene L.
L. Francis
Germain L.
Long L.
Clear L.
Big L.
Hard L.
Rice L.
Grass L.
Ryan L.
Elizabeth L.
Typha L.
Bass Bluff L.
Red till
Outlet Brook
Sand plain
Gray till
Red gravel and till
Mississippi Valley train
Gray till
Sand plain
Gray till
Gray till
Gray till
Gray till
Sand plain
Sand plain
Gray till
Gray till
Gray till
Gray till



LAKE OF THE WOODS, SOUTHWESTERN PORTION, MINNESOTA AND ONTARIO

Shown are surficial and bedrock geology, bottom sediment types, sieve analyses, moisture content, and pH and Eh values of wet core samples. Abbreviations: p, testate protozoans; c, cladocerans; d, diatoms, dm, abundant melosiroid diatoms; dc, abundant coscinodiscoid diatoms; o, ostracods; e, egg cases; w, pondweed fragments; abundant where underlined

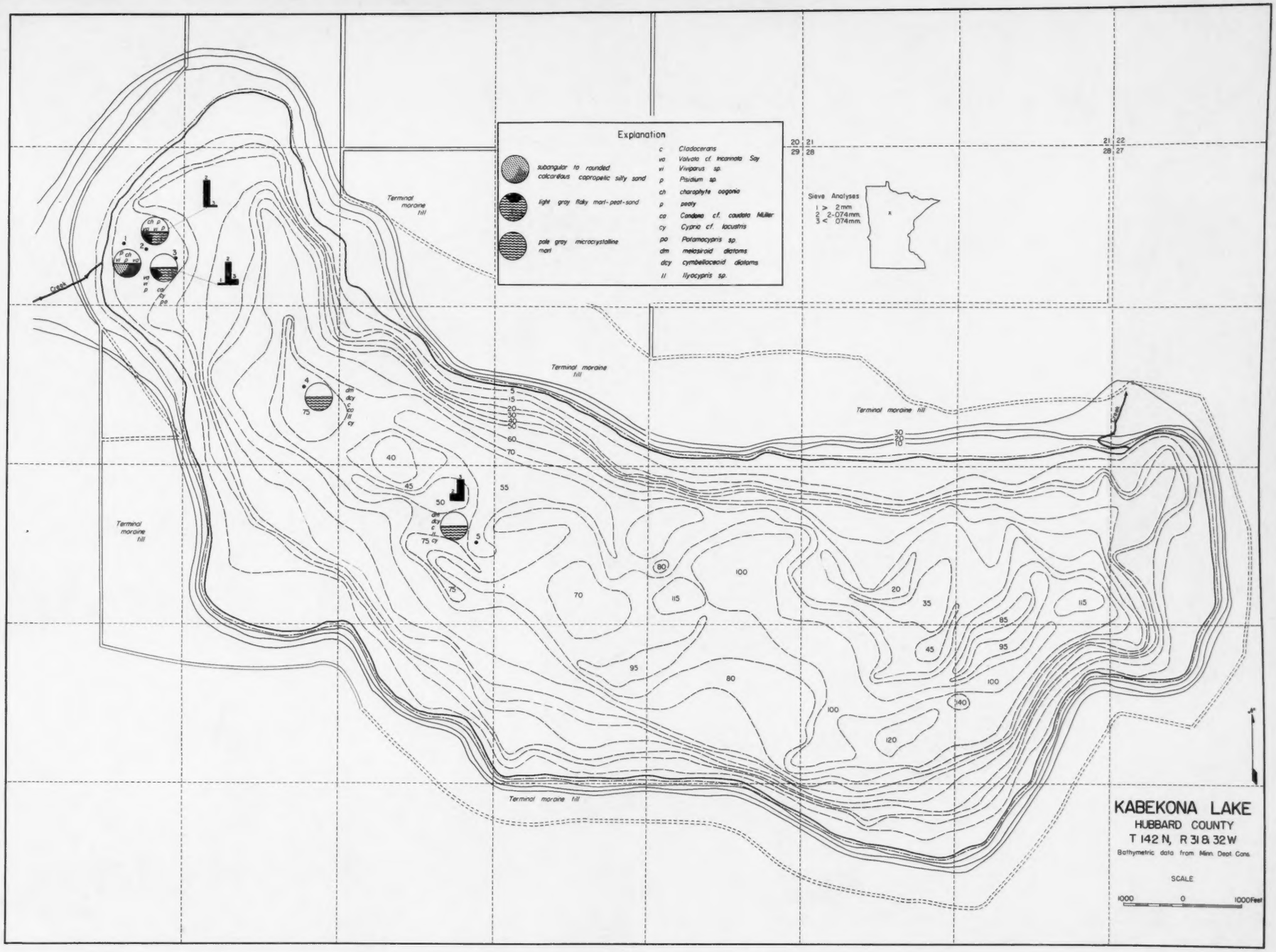
SWAIN, PLATE 2
Geological Society of America Bulletin, volume 72



LAKE OF THE WOODS, SOUTHWESTERN PORTION, MINNESOTA AND ONTARIO

Shown are surficial and bedrock geology, bottom sediment types, sieve analyses, moisture content, and pH and Eh values of wet core samples. Abbreviations: p, testate protozoans; c, cladocerans; d, diatoms, dm, abundant melosiroid diatoms; dc, abundant coscinodiscoid diatoms; o, ostracods; e, egg cases; w, pondweed fragments; abundant where underlined

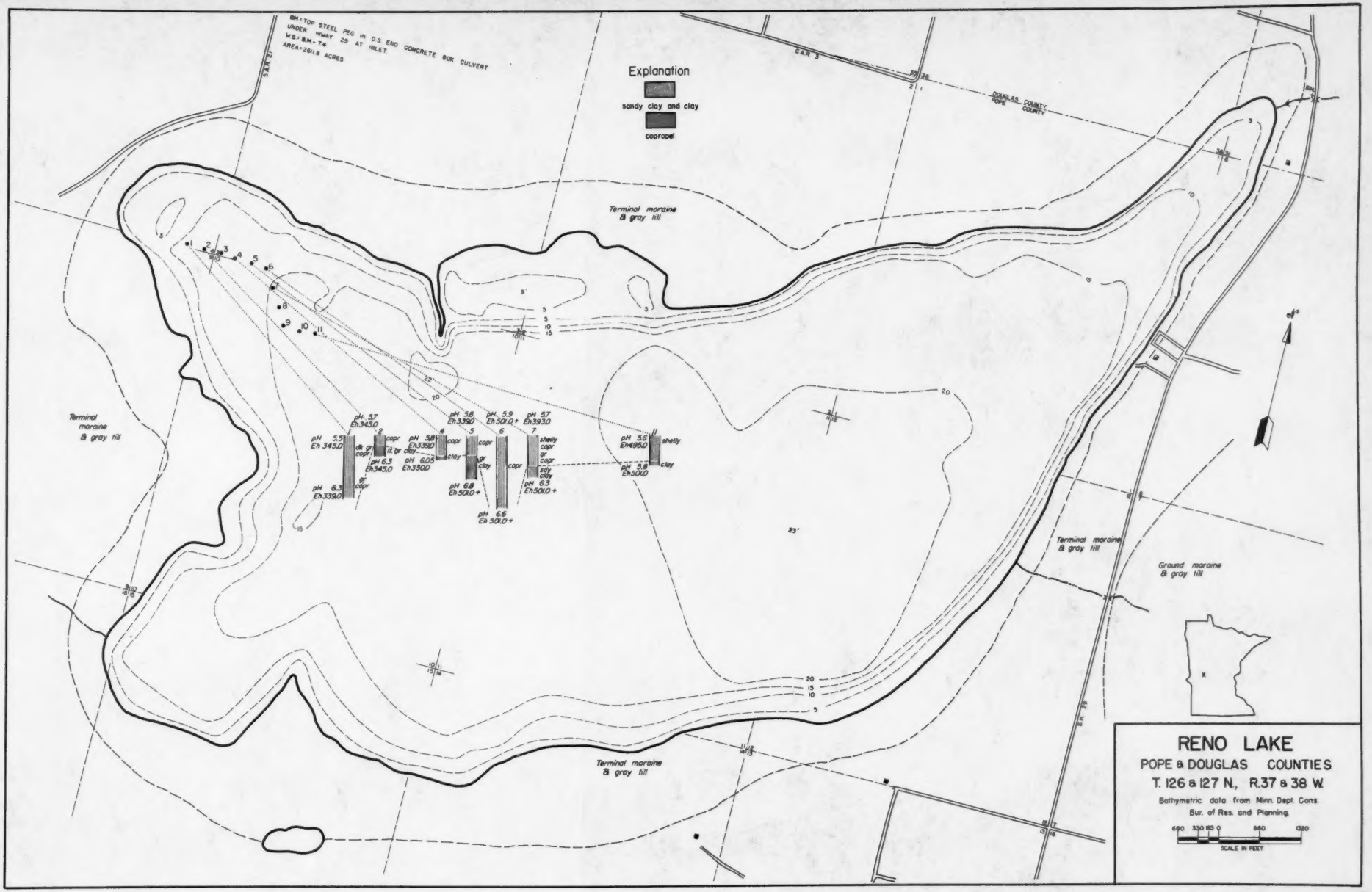
SWAIN, PLATE 2
Geological Society of America Bulletin, volume 72



KABEKONA LAKE, HUBBARD COUNTY, MINNESOTA

Shown are surficial geology, types of bottom sediments, and sieve analyses of sediments. Abbreviations are listed on map

SWAIN, PLATE 3
Geological Society of America Bulletin, volume 72


BM: TOP STEEL PEG IN D.S. END CONCRETE BOX CULVERT UNDER HWAY 29 AT INLET
W.S. B.M.-74
AREA-2618 ACRES
Explanation
sandy clay and clay
copropel
Terminal moraine & gray till
Ground moraine & gray till
DOUGLAS COUNTY
POPE COUNTY
C.A.R. 3
S.A.R. 27
T.H. 29
RENO LAKE
POPE & DOUGLAS COUNTIES
T. 126 & 127 N., R. 37 & 38 W.
Bathymetric data from Minn. Dept. Cons. Bur. of Res. and Planning.
SCALE IN FEET

RENO LAKE, POPE AND DOUGLAS COUNTIES, MINNESOTA

Shown are surficial geology, types of bottom sediments, pH and Eh of sediments. Eh values are in mv

SWAIN, PLATE 4
Geological Society of America Bulletin, volume 72

SHOREHAM
WHITING
LEICESTER
ORWELL
BRANDON
ADDISON CO.
RUTLAND CO.
Sunset Lake
Lake Dorothea
Lake Hortonia
Burr Pond
Hinkum Pond
Beebe Pond
Austin Pond
Mill Pond
Sudbury
Brandon
Forestdale
Leicester Junction
Whiting River
Otter Creek
Otter Brook
Giddings Brook
Kehler Pond Gauge
Fault
A
A'
B
B'
C
C'
D
D'
E
E'
F
F'
G
G'
H
H'

D E F G H I J K L M N

0 1 2 3 4 5 6 7 8

EXPLANATION

SYNCLINORIUM SEQUENCE

Ira formation
(East side of Taconic Sequence)

Oi; *black, siliceous to graphitic phyllite, locally limy or with beds rich in albite porphyroblasts and dark to greenish grey in color.*

Oiw; *Whipple marble member: dark bluish grey-weathering dark grey marble, generally finer-grained than the Lower Ordovician carbonates. Basal to the Ira formation, but locally interbedded with the lower strata of the slates and phyllites.*

Hortonville formation
(West side of Taconic Sequence)

Oh; *black, siliceous to graphitic slate, locally limy or silty and weathering reddish.*

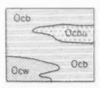
Ohh; *Forbes Hill conglomerate: angular and poorly sorted quartzite-and-slate pebbles in black slate matrix.*

Ohl; *coarse limestone-pebble, limy-matrix conglomerate in black slate.*

Undifferentiated Mid-Ordovician limestones
(Middlebury, Orwell, and Glens Falls)

Otc; *Undifferentiated dark grey, thin-to-medium bedded limestone and marble, with local interbedding of thin brown-weathering dolomitic layers. Local siliceous and black phyllitic partings in the limestone.*

MIDDLE ORDOVICIAN



Chipman formation

Ocb; *Beldens member; grey, massive marble with interbedded massive creamy dolostone; white massive marble with red hematite streaks.*

Ocbu; *Burchards member: similar to Beldens member but characterized by buff-weathering dolomitic mottles in limestone and marble beds.*

Ocw; *Weybridge member: grey fine marble interbedded with brown-weathering, slightly dolomitic layers at equal intervals of about 3 inches. Locally these contrasting strata show cross-bedding.*

Bascom formation

Obc; *grey, streaky marble; grey marble with dolomitic mottles; massive white marble; interbedded massive grey dolostone and grey marble; interbedded thin grey-white marble and brown-weathering siliceous dolostone and limestone; massive spongy-weathering dark grey dolomitic quartzite; grey sugary marble with pale siliceous or micaceous partings.*

LOWER ORDOVICIAN

TACONIC SEQUENCE

Pawlet formation

Opw; *silver grey to jet black, locally graphitic and pyritiferous, micaceous silty slate interbedded with grey, locally calcareous greywacke. The greywacke weathers dull brown and commonly shows graded bedding. Grains of subangular quartz, feldspar, and rock fragments up to 1/8 inch in a calcareous and argillaceous matrix. Beds up to 4 feet thick. Basal part of the formation carries graptolites.*

MIDDLE ORDOVICIAN

Mount Hamilton Group

O-Cmh; *black, grey, green, purple, to red hard slates, weathering white (units 1, 2, and 3 of the text); with beds of ankeritic quartzite from a few inches to several feet thick (unit 5, bright red); latter may contain layers of an edgewise conglomerate. Some of the slates are interbedded with thin cherty quartzites or ribbon limestones a few inches apart. A red, smooth, soft slate is included (unit 4), as well as a lime-matrix polymikt limestone conglomerate (unit 6), including black chert pebbles. Lithology of the group varies rapidly along strike and separation from Cwc is not always easy. Abundant cross-bedding, channel-filling, etc.*

UPPER CAMBRIAN TO MIDDLE ORDOVICIAN

West Castleton formation

Cwc; *grey siliceous to black, graphitic, pyritiferous slate and phyllite, locally with interbedded thin dark grey dolostone and grey quartzite and arkosic layers. Thin, white sandy laminae common in the graphitic variety.*

Cwcb; *Beebe limestone member: massive, lenticular black limestone weathering blue-grey. Fossiliferous.*

Bull formation

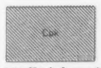
Cbmt; *Mettawee slate facies: bulk of the formation. Purple, green-grey, and variegated slate and phyllite, non-chloritoid bearing; minor beds of white to green quartzite and creamy limestone near the top of the formation.*

Bright blue; *North Brittain conglomerate member: limestone-pebble conglomerate in a non-calcareous green to purple slate matrix. Thickness extremely variable. The conglomerate is intraformational and may pass visibly into boudinaged limestone beds. Fossiliferous. Near West Castleton village, the unit is a grey ribbon-limestone in a dark grey slate matrix.*

Bright green; *Mudd Pond quartzite: white to grey, vitreous, medium-grained orthoquartzite with local dolomitic pods (concretions?). Locally dark grey quartz-grit in a slaty matrix (Eddy Hill lithology). Rock generally weathers white and smooth, with a waxy luster.*

Cbzh; *Zion Hill member: green, vitreous, chloritic quartzite or greywacke with common limonite spots. Base of the unit is commonly a pebble conglomerate and top may be a siltstone. Unit as a whole is massive and lenses in and out rapidly. Weathers white and tends to form cliffs. East of Lake Bomoseen, it is stratigraphically below the Bomoseen greywacke, but west of the lake, it is above the Bomoseen.*

Cbbm; *Bomoseen greywacke member: green to olive-colored arkose and greywacke, carrying visible flakes of mica and rock fragments, and microscopic stilpnomelane. Weathers pale red to white. Massive and uniform, locally with white quartzite layers.*



Biddie Knob formation

Cbk; *purple and green chloritoid-bearing slate and phyllite. Minor beds of white to green quartzite and rarely limestone seams. Especially south of Castleton River but also locally north thereof, the Zion Hill quartzite is actually within, though near the top of, this formation.*

LOWER CAMBRIAN

CONTACTS

Sedimentary contact
Dashed where inferred or interpolated, dotted where concealed. Queried

Base from U.S. Geological Survey 7½-minute topographic sheets: Benson (1946), Bomoseen (1944), Brandon (1946), Orwell (1949), Proctor (1944), Sudbury (1946), and Thorn Hill (1946), U.S.G.S. 15-minute topographic sheet: Castleton (1895), U.S. Soil Conservation Service aerial photographs, DCC series, 1941-42.

Scale of Castleton sheet 1:62,500; Scale of other sheets 1:24,000

Williams & Heintz Map Corporation Washington 27, D. C.

Geology by E-an Zen, assisted by W. D. Means, A. J. Murray, and D. Harwood, 1953-58.

BEDROCK GEOLOGY OF THE CASTLETON AREA, VERMONT

Scale 1:48 000

1 0 1 2 Miles

ZEN, PLATE 1

Geological Society of America Bulletin, volume 72

Biddie Knob formation

Cbk: *purple and green chloritoid-bearing slate and phyllite. Minor beds of white to green quartzite and rarely limestone seams. Especially south of Castleton River but also locally north thereof, the Zion Hill quartzite is actually within, though near the top of, this formation.*

CONTACTS

Sedimentary contact

Dashed where inferred or interpolated, dotted where concealed. Queried where interpreted.

Fault contact

Generally low-angle reverse (thrust) faults. T on upper plate. Dashed where inferred or interpolated, dotted where concealed. Queried where interpreted.

*

Fossil locality

CF (Z): *Cambrian locality; this report.*

CF (S): *Cambrian locality; reported by Swinnerton and Schuchert.*

CF (F): *Cambrian locality; reported by Fowler*

CF (D): *Cambrian locality; reported by Dale and assistants (notebooks)*

OF (D): *Ordovician locality; reported by Dale and assistants (notebooks)*

Williams & Heintz Map Corporation Washington 27, D. C.

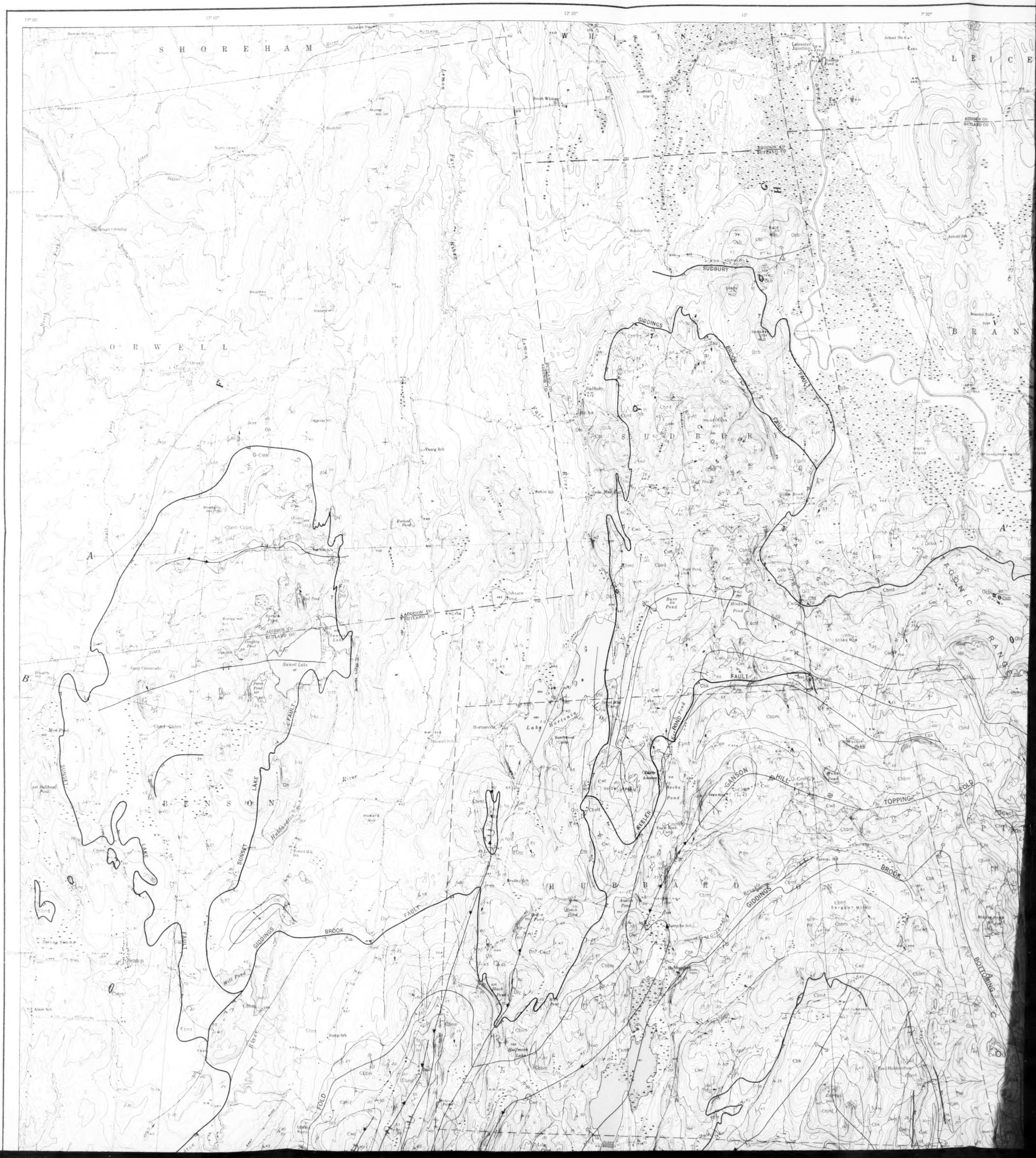
Geology by E-an Zen, assisted by W. D. Means, A. J. Murray, and D. Harwood, 1953-58.

BEDROCK GEOLOGY OF THE CASTLETON AREA, VERMONT

Scale 1:48 000

1 0 1 2 Miles

n, volume 72


SHOREHAM
WHITING
LEICE
ORWELL
SUDBURY
BRAN
BENSON
HUBBARDTON
ADDISON CO
RUTLAND CO
Lemon Fair River
Fox Island
Arnold Brook
Pepper Brook
Mill Brook
Otter Creek
Lake Dunmore
Lake Hortonia
Sunset Lake
Mud Pond
Burr Pond
Hinkum Pond
Halfmoon Lake
Black Pond
Roaring Swamp
Barnums Hill
Hardigan Hill
Murray Hill
Woodchuck Hill
Slaty Hill
Stiles Hill
Howard Hill
Green Mtn
Zierings Hill
Leicester Junction
South Whiting
Whiting
Sudbury
Orwell
Benson
Hubbardton
Brandon State Sch
Halls Island
Beaman Island
Webster Sch
Ferton Sch
Turnpike Sch
Aiken Sch
Bump Sch
Bradley Sch
Hortonville
Beechwood Camp
Knox Rock Camp
Lilac Pond
Sargent Marsh
Parsons Sch
Walker Pond
Arnold Sch
School No 4
Thrush Crossing
Stacy Crossroads
Daigneau Mill
Hibbert Hill
Howard Hill Sch
Thingall Sch
Yaung Sch
Perkins Pond
Perch Pond
Spruce Pond
Doughty Pond
Johnson Pond
Rocky Pond
Beebe Pond
Austin Pond
Walker Pond
Mud Pond
Halfmoon Lake
Lily Pond
Sunset Lake
Lake Dunmore
Lake Hortonia
Mill Pond
Roaring Swamp
Burr Pond
Hinkum Pond
Stiles Hill
Zierings Hill
Green Mtn
Slaty Hill
Woodchuck Hill
Murray Hill
Barnums Hill
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Hortonville
Beechwood Camp
Knox Rock Camp
Sargent Marsh
Parsons Sch
Walker Pond
Brandon State Sch
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Slaty Hill
Stiles Hill
Green Mtn
Zierings Hill
Leicester Junction
South Whiting
Whiting
Sudbury
Orwell
Benson
Hubbardton
Brandon State Sch

EXPLANATION

CONTACTS

Normal sedimentary contact
Degree of control not specified

Fault contact; commonly low-angle reverse fault
T *on the side of the upper plate. Degree of control not specified*

STRUCTURE SYMBOLS

Note: All data recorded and plotted to the nearest five degrees

50
Strike and dip of bedding
Including inverted beds

Strike of vertical bedding

Horizontal bedding

70
Strike and dip of cleavage

Strike of vertical cleavage

30
Strike and dip of bedding with parallel cleavage

Strike of vertical bedding and parallel cleavage

15
20
Bedding and cleavage with parallel strike
First figure gives the dip of the bedding; second figure gives the dip of the cleavage

30
55
Bedding and cleavage with parallel strike, but dip in opposite directions

40
25
Bedding and cleavage not parallel
Point of junction of the truncated symbols represents the location of the outcrop

60
30
Two cleavages with parallel strikes, but different dips
First figure refers to (a) the earlier cleavage, if the sequence is determinable; (b) the better formed of the two cleavages, including foliation in the phyllites; or (c) the more closely-spaced of the two cleavages.

25
70
Two non-parallel cleavages from the same outcrop

45
10
Axial plane of fold
Arrow gives the direction and amount of the true plunge as measured in a vertical plane

65
Axial plane of fold with no plunge

35
Vertical axial plane
Arrow gives sense and amount of the plunge

10
Lineation, with sense and amount of the true plunge as measured in a vertical plane
Butt-end of arrow indicates location of outcrop

45
35
Bedding with lineation
Arrow gives direction and amount of the plunge

80
65
Cleavage with lineation
Arrow gives direction and amount of the plunge

30
Strike and dip of exposed fault surface

SYMBOLS PERTAINING TO AXIAL SURFACES

Trace of axial surface on the topography
Arrow indicates the projected sense of the axial plunge

Axial trends of simple anticlines

Axial trends of simple synclines

Axial trends of simply overturned anticlines

Axial trends of simply overturned synclines

Trace of axial surface of bottoming fold on the topography
Arrow gives approximate direction and sense of plunge of the axis of the fold at the particular locality, and barb is on that side of the arrow in the direction of facing

Trace of axial surface of topping fold on the topography
Arrow gives approximate direction and sense of plunge of the axis of the fold at the particular locality, and barb is on that side of the arrow in the direction of facing

SYMBOLS FOR THE FORMATIONS

WEST HAVEN
FAIR HAVEN
HAMPTON
WHITEHALL
CASTLETON
HUBBARDTON
POULTNEY
MIDDLETOWN
IRA
Point of Pines
Castleton Corners
Castleton
Hydeville
Parker Hill
Fair Haven
Glissville
Pond Hill
Bird Mountain
Clark Hill
Gorhamtown
Herrick Mountain
Old Knob
Spruce Knob
Spaulding Hill
Poultney
East Poultney
Town Hill
Graham Hill
Moss Hollow
Clark Hollow
Hooker Hill
Eaton Hill
Barker Hill
Blueberry Hill
Cedar Mtn
Prospect Hill
Ben Hill
Zion Hill
Mason Pt
Crystal Lake
Hillis Pond
Lake Pond
Melvin Pond
Prowse Pond
Gully Brook
Vail Brook
Avorn Brook
Train Brook
Castleton River
Hudson and Castleton R.R.
Delaware and Hudson R.R.
Rutland and Washington R.R.
Hubbardton
Great Ledge
Bottoming Ridge
Porcupine Ridge
Scotch Hill
Mt Hamilton
Pine Hill
Graham Hill
Hooker Hill
Eaton Hill
Barker Hill
Blueberry Hill
Zion Hill
Cedar Mtn
Prospect Hill
Ben

Axial trends of simply overturned anticlines

Axial trends of simply overturned synclines

Trace of axial surface of bottoming fold on the topography
Arrow gives approximate direction and sense of plunge of the axis of the fold at the particular locality, and barb is on that side of the arrow in the direction of facing

Trace of axial surface of topping fold on the topography
Arrow gives approximate direction and sense of plunge of the axis of the fold at the particular locality, and barb is on that side of the arrow in the direction of facing

SYMBOLS FOR THE FORMATIONS

- Oi: Black slate and phyllite of the Ira formation.
- Oiw: Whipple marble member of the Ira formation.
- Oh: Black slate of the Hortonville formation, including the Forbes Hill conglomerate.
- Otc: Undifferentiated Chazy, Black River, and Trenton limestone.
- Ocb: Beldens member of the Chipman formation.
- Ocbu: Burehards member of the Chipman formation.
- Ocw: Weybridge member of the Chipman formation.
- Opw: Pawlet formation.
- O-Cmh: Mount Hamilton group; figures following the symbol indicates lithologic units.
- Cwc: West Castleton formation.
- Cbmt: Mettawee facies of the Bull formation, but including the North Brittain conglomerate member, the Mudd Pond quartzite member, and the Zion Hill member.
- Cbbm: Bomoseen member of the Bull formation.
- Cbk: Biddie Knob formation.

STRUCTURAL DATA OF THE BEDROCK GEOLOGY, CASTLETON AREA, VERMONT

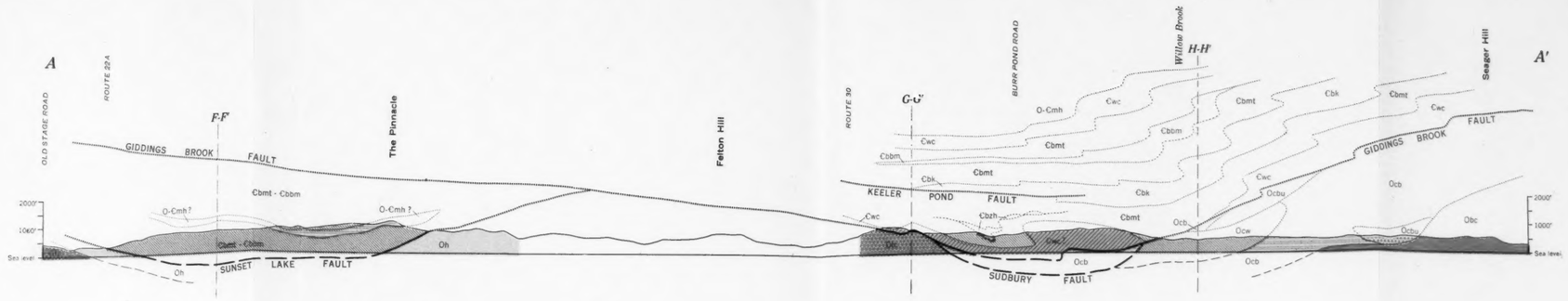
Scale 1:48 000

1 0 1 2 3 4 5 Miles

Williams & Heintz Map Corporation Washington 27, D. C.

S 78° W

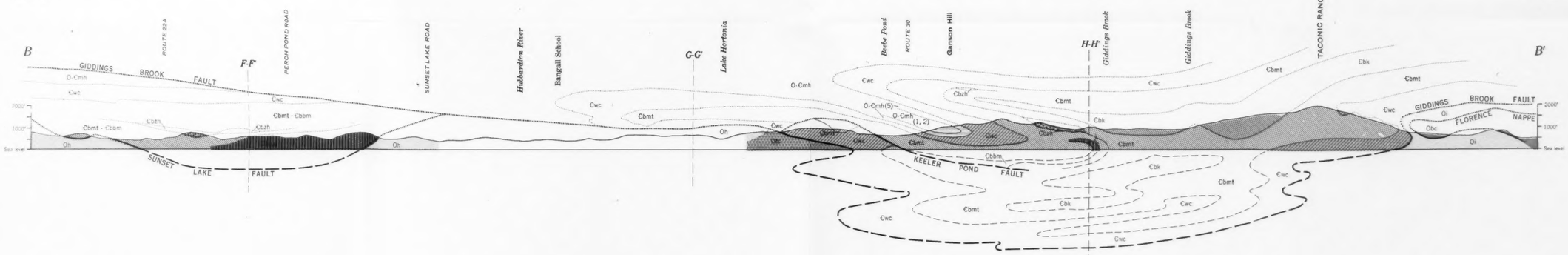
N 78° E



N 5° E

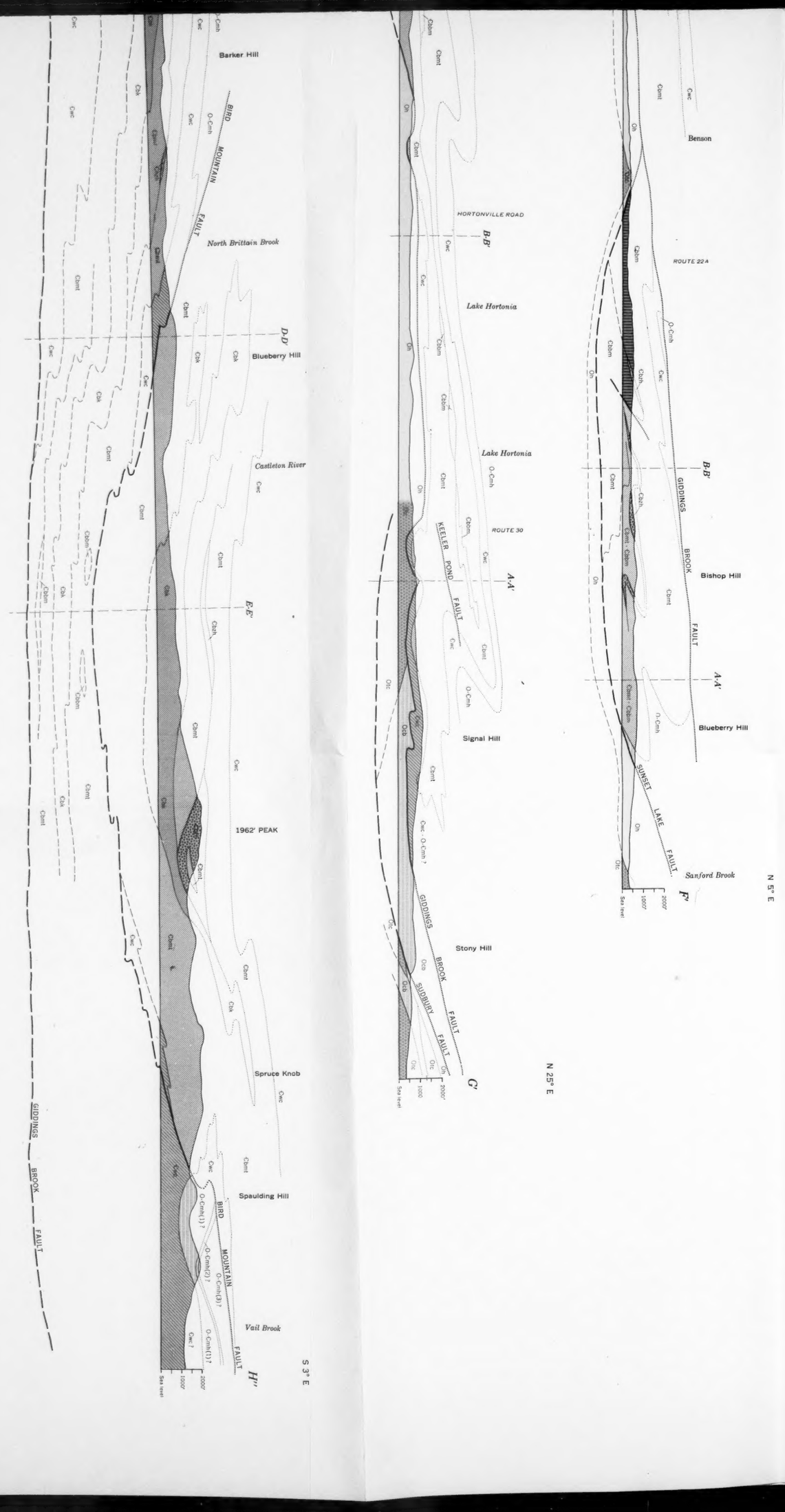
N 78° W

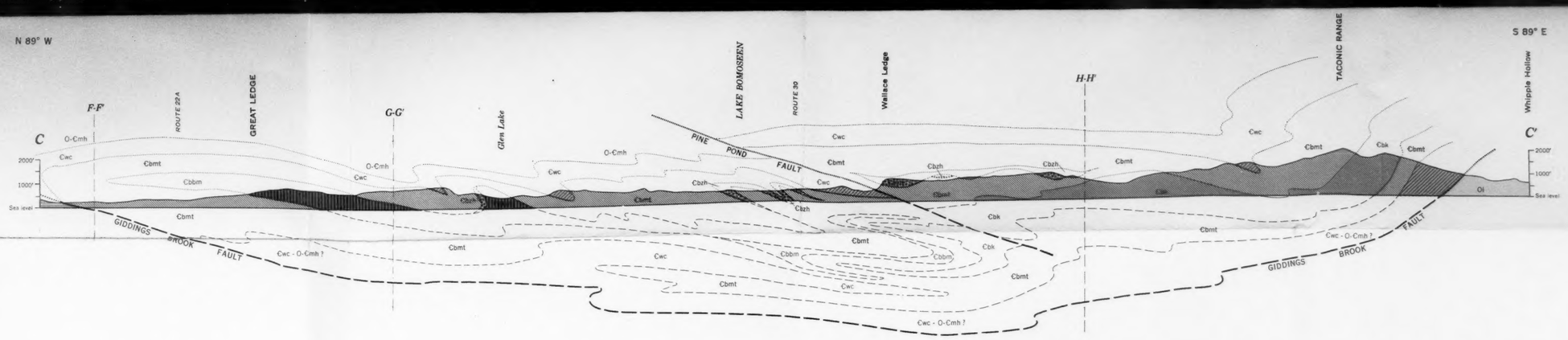
S 78° E

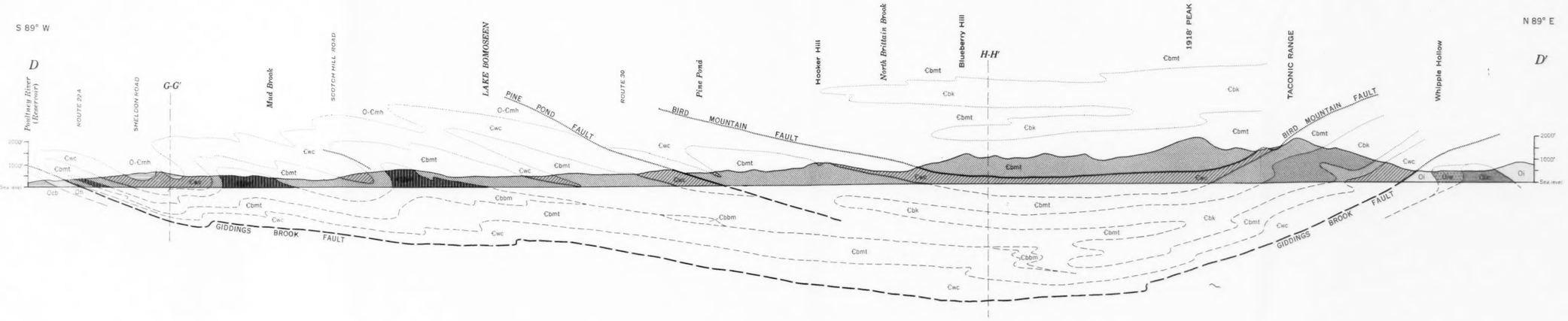


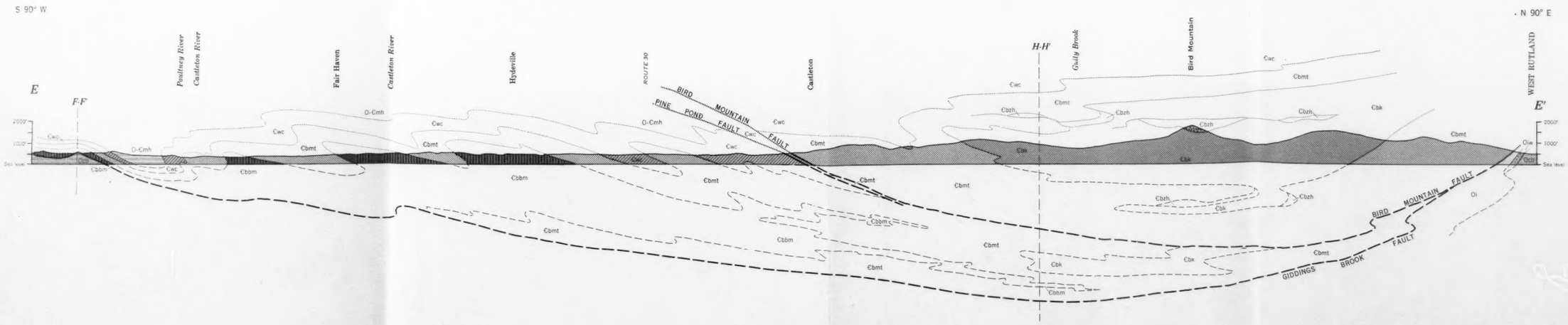
N 89° W

S 89° E


Barker Hill
BIRD MOUNTAIN FAULT
North Brittain Brook
D-D'
Blueberry Hill
Castleton River
E-E'
1962' PEAK
Spruce Knob
GIDDINGS BROOK FAULT
Spaulding Hill
Vail Brook
H-H'
S 3° E
HORTONVILLE ROAD
B-B'
Lake Hortonia
ROUTE 22A
Benson
Lake Hortonia
ROUTE 30
KEELER POND FAULT
A-A'
Signal Hill
Stony Hill
GIDDINGS BROOK FAULT
SUDBURY FAULT
G-G'
N 25° E
Bishop Hill
GIDDINGS BROOK FAULT
Blueberry Hill
SUNSET LAKE FAULT
Sanford Brook
F-F'
N 5° E
Whipple Hollow
C-C'
S 89° E
TACONIC RANGE
H-H'
Wallace Ledge
FAULT
KEELER POND FAULT
GIDDINGS BROOK FAULT
NAPPE
FLORENCE
B-B'
S 78° E
TACONIC RANGE
H-H'
Giddings Brook
Giddings Brook
Ganson Hill
ROUTE 30
Bridge Pond
SUDBURY FAULT
KEELER POND FAULT
GIDDINGS BROOK FAULT
G-G'
ROUTE 30
A-A'
Seager Hill
H-H'
Willow Brook
BURR POND ROAD
N 78° E





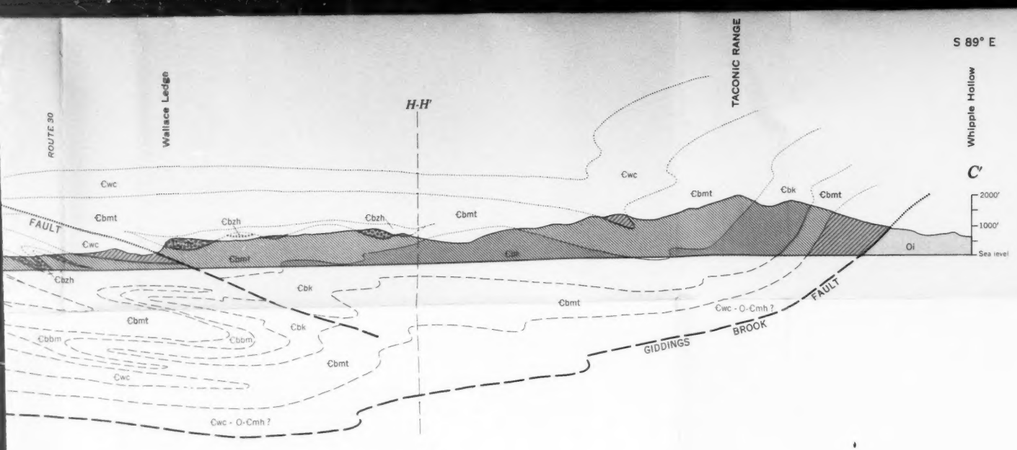


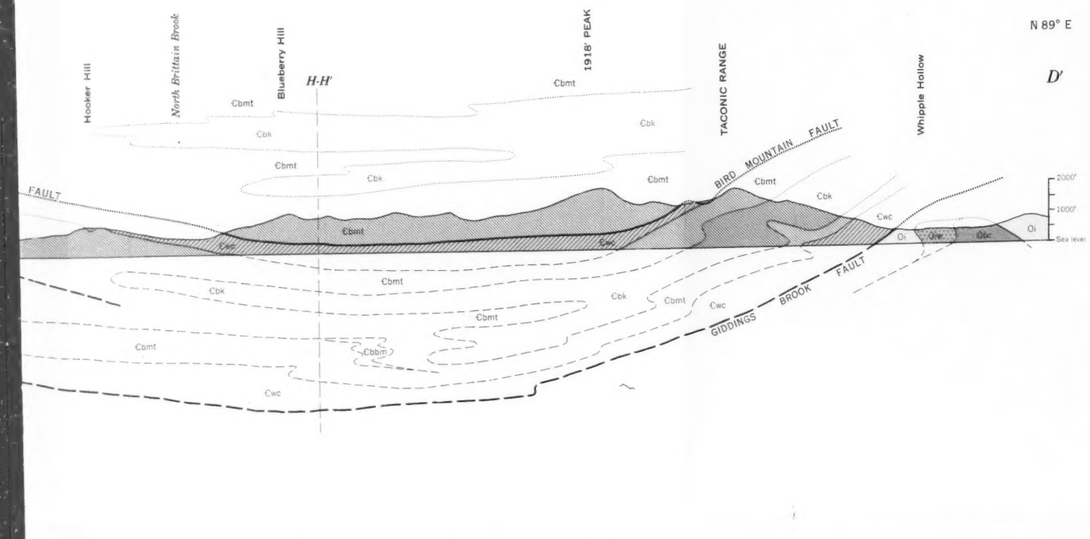
GEOLOGIC SECTIONS ACROSS THE CASTLETON AREA, VERMONT

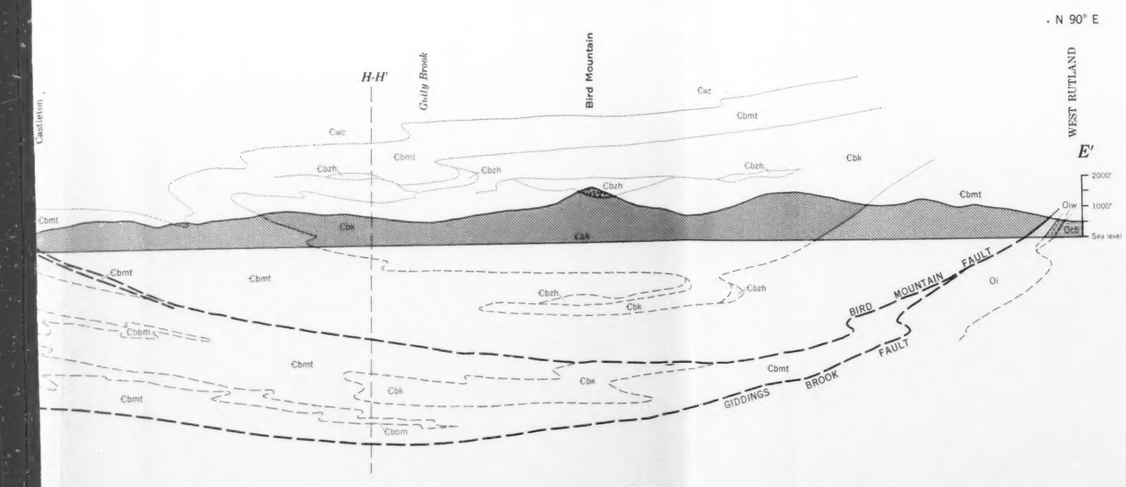
Scale 1:48 000

1 0 1 2 3 4 5 Miles

ZEN, PLATE 5
Geological Society of America Bulletin, volume 72

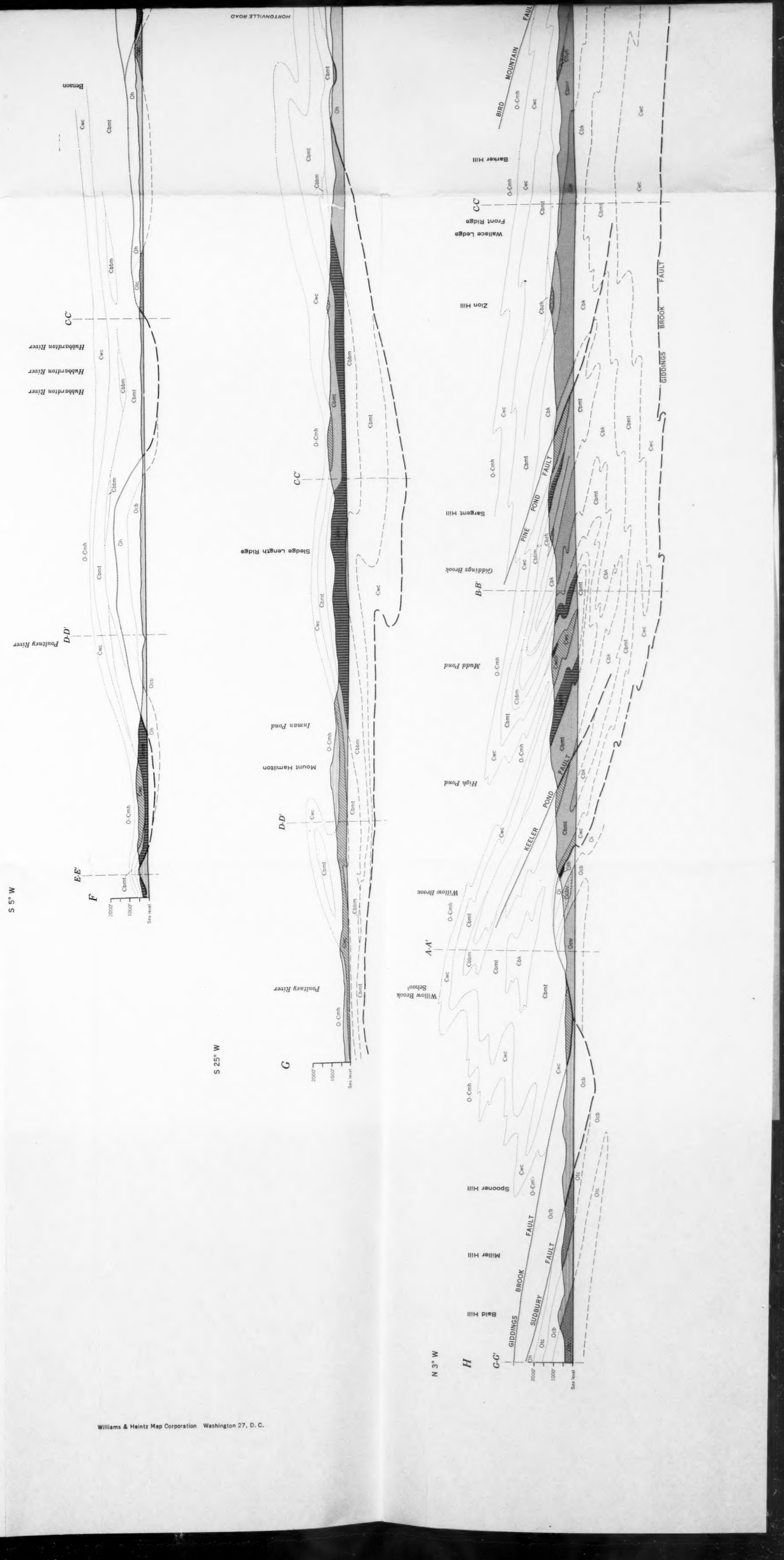






E CASTLETON AREA, VERMONT

000 3 4 5 Miles



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